

Precambrian geology
of the Västana area,
southern Sweden

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PRECAMBRIAN GEOLOGY
OF THE VÄSTANÅ AREA
SOUTHERN SWEDEN

av

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with one map

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CONTENTS

ABSTRACT	4
1. INTRODUCTION	5
2. PREVIOUS WORK	5
3. METHODS OF INVESTIGATION	6
4. PETROGRAPHIC DESCRIPTIONS	7
Metavolcanic formation	7
Metasedimentary formation	29
Gneiss complex	41
Granitoid rocks	50
Dolerite	52
5. STRUCTURAL HISTORY	52
Introduction	52
Planar structures	53
Linear structures	55
Boudinage	56
D ₁ -deformation	59
D ₂ -deformation	60
D ₃ -deformation	63
D ₄ -deformation	64
Tentative structural development	65
6. TIME RELATION OF DEFORMATION AND MINERAL GROWTH	66
7. METAMORPHIC HISTORY	72
8. GENERAL CONSIDERATIONS	76
9. ACKNOWLEDGEMENTS	79
10. REFERENCES	80

ABSTRACT

The present paper deals with the petrology and tectonics of the Precambrian of the Västana area, southern Sweden. The area is predominantly made up of volcanics and sedimentary rocks, which are intruded by a younger granitoid. Mineralogical and chemical data are given for the various rock-types. Particular attention is given to the genesis and structural history. Several periods of deformation have been distinguished. The supracrustal rocks are considered to be of Svecofennian age, whereas the younger granitoids are post-Svecofennian.

1. INTRODUCTION

This work describes the geology in the Västana area, which is situated in the northeastern part of Skåne, southernmost Sweden (see inset).

In the eastern part of the area there occurs a gneiss (Coastal Gneiss), which passes into porphyritic tuffs. Sediments make up the central part of the area and the western part comprises fine-grained gneisses. The southern and northeastern parts are occupied by granitoids.

The area attracts geological interest due to its proximity to the belt of tectonization separating southeastern and southwestern Sweden (the Småland Suture or Protogine Zone) and due to the presence of rather well preserved supracrustals, which contain coexisting andalusite and kyanite. Another subject of much discussion has been the relation between the Coastal Gneiss and the porphyritic tuffs.

The aim of the present investigation has been to study the petrology, tectonics and the relative age relations of the different rocks. Much emphasis has been put on the genesis and the structural history. However, because of the complicated development of the Precambrian rocks in the region and the lack of isotopic data the outlines of the geological history suggested in the final chapter are very tentative.

2. PREVIOUS WORK

The first regional surveys, which included the Västana area, were carried out by G. de Geer (1889) and A. Blomberg (1892). Outlining the main geological features these pioneers reported that the Precambrian rocks in the area included gneisses, mica-quartzites, conglomerates, dioritic schists, granites, pegmatites, and dolerites.

Later, in 1897, a detailed study of the Precambrian rocks in the Västana area was made by Bäckström. This work gives an excellent petrographic description of the rock-types. Thus Bäckström describes

a sequence of rocks, which includes, commencing with the youngest, granites, quartzite, amphibolites, dense fine-grained gneiss, which passes into a fine-grained gneiss (Coastal Gneiss), and the magnetite-gneiss formation.

The genesis of the Coastal Gneiss has been the subject of several investigations. The most extensive and detailed papers have been published by R. Norin (1936 and 1959). In the 1959 publication he sustained the observation made by Bäckström, that no sharp boundary exists between the foliated tuffs and the Coastal Gneiss, and thus considered that the Coastal Gneiss is a strongly metamorphosed equivalent of the quartz-porphyrific tuffs. This view differs from the opinion of Asklund (1947), who regarded the Coastal Gneiss as a variety of the "gneiss-granites".

According to Norin (1959) the geological evolution of Blekinge and eastern Skåne may be summarized as follows:

1. Deposition of supracrustals (Västana complex).
2. Oldest orogenesis. Plutonic activity and deformation of the supracrustals.
3. Volcanic and plutonic activity.
4. Youngest orogenesis. Plutonic activity and migmatization.
5. Post-orogenic plutonic activity.
6. Diabas intrusions.

Among the other contributors to the regional geology of western Blekinge and eastern Skåne Moberg (1896), Habetha (1936) and Zeck (1971) merit mention.

3. METHODS OF INVESTIGATION

For the field work the Topographical Map of Sweden (scale 1:50 000) and aerial photographs (scale 1:20 000) have been used. The aerial photographs have also been used in the photogeological interpretation of the fault systems.

For a reference to observation and sampling localities the Universal Transverse Mercator system (UTM, with an accuracy of ± 100 m) is used. The petrological map (see inset) is based on

studies of outcrops and interpretation of magnetic maps, which have kindly been put to the author's disposal by the Geological Survey of Sweden (SGU).

The feldspar determinations have been made by universal stage employing the zone method for the plagioclases. In some cases the position of the extinction in relation to cleavage and twinning were also determined.

Refractive indices were determined in monochromatic (Na_D) light by the immersion method.

The modal analyses were carried out on stained thin sections and the results were calculated according to the methods of Van der Plas and Tobi (1965). The data are given as per cent by volume. For the semi-quantitative description of mineral compositions the following nomenclature has been used:

- >25 % Main minerals.
- 5-25 % Essential minerals.
- 1-5 % Subordinate minerals.
- <1 % Accessory minerals.

Whole-rock analyses were made at the laboratories of the Geological Survey of Sweden and the Geological Institute, Lund. The sample weights were between 2 and 4 kilograms.

Most petrochemical calculations were made by the SGU Data-section using the Benorm Program. Other calculations of rock analyses are according to Burri (1959).

4. PETROGRAPHIC DESCRIPTIONS

4.1 Metavolcanic formation

Acid metavolcanics

Acid, quartz- and feldspar-porphyrific metavolcanics constitute the bulk of the area marked as metavolcanics on the map. They form together with small layers of mica-schists, meta-arenites and tuffites the basal part of the supracrustal group and extend from

Blistorpasjön in the south-east to the middle part of Lake Immeln (see map). The acid metavolcanics are light to medium grey. They are almost always porphyritic with phenocrysts of plagioclase and to a lesser extent also of quartz. The phenocrysts never exceed 20 % of the rock volume.

Distinct layers of mica-schists and meta-arenites are most frequent in the upper parts of the acid metavolcanic formation. Less well defined are the thin tuffite units. They are transitional between metavolcanics and metasediments. Phenocrysts are also present in the tuffites. Banded metavolcanics occur locally and ran gradually into massive rock-types. The foliation is normally distinct, but may in some outcrops be diffuse or absent.

The texture varies from predominant granoblastic-polygonal to more or less xenoblastic. The matrix is completely recrystallized, and apart from the phenocrysts no other microscopic primary features have been observed. The bulk of the groundmass is made up of quartz, plagioclase and alkali feldspar. Biotite, and to a lesser extent muscovite, occur in essential or subordinate amounts. Accessories are chlorite, sphene, leucoxene, zoisite, allanite, calcite, rutile, zircon, hematite, and magnetite.

The alkali feldspar is usually (string- and flame-) perthitic and the plagioclase is albitic. Quartz makes up a major part of the groundmass and is often weakly undulatory. At the grain boundaries there are small scattered flakes of biotite and muscovite. Green biotite occurs in flakes elongated parallel with the foliation. Alteration products such as leucoxene often appear in the biotite. Pleochroic halos are frequent. At or close to the contact with secondary quartz veins the biotite is partly replaced by chlorite. Muscovite occurs at the grain boundaries and as alteration product in the phenocrysts. Larger crystals are often poikilitic and randomly orientated. Intergrowths of muscovite and biotite appear sporadically. Epidote is a subordinate mineral and is sometimes twinned. Cores of allanite in the epidote are common.

The plagioclase phenocrysts never exceed 20 % of the rock volume and the grain-size is never larger than 4 mm. Remnants of plagioclase, which often are only skelton (Fig.2, p.10), are more or less sericitized and contain small inclusions of fresh feldspar

and quartz. The phenocrysts are commonly enveloped by clear rims, which comprise albite, perthitic alkali feldspar and quartz. A more complete destruction of the plagioclase may lead to a poikilitic host of alkali feldspar, which encloses minute remnants of the plagioclase phenocrysts. Normally the phenocrysts are strongly granulated. In these cases the phenocrysts can only be traced as assemblages of somewhat coarser grain-size and from the pattern of parallel texture, which is accentuated by mica flakes, and swirls around these aggregates (Fig. 4). The An-content of the plagioclase varies between 26 and 34 %. Thus the plagioclase phenocrysts are more basic than the plagioclase individuals in the matrix or the secondary alteration products.

The quartz phenocrysts (Fig. 3) are in general irregularly rounded and subdivided into individuals of different crystallographic orientation. Grains with strongly accentuated undulatory extinction have sutured grain boundaries.

The acid metavolcanics are rather homogeneous. Nevertheless, there exist differences in the contents of phenocrysts, colour, banding, etc. Below, some of these modifications of acid metavolcanics will be described in detail.

North of N. Grönhult (643/338), there outcrops a medium grey, slightly foliated metavolcanic. The plagioclase phenocrysts are subrounded and have maximum sizes of 15 mm. They are light greenish-grey in colour. Most often the phenocrysts are assembled into lenticular aggregates, which are enveloped by thin, sparse biotite flakes. Recrystallized alkali feldspar, perthite and quartz are not uncommon in pressure shadow zones. Lens-shaped assemblages of mafic minerals are frequent in connection with the phenocrysts. As in the case with the phenocrysts, these aggregates are largely parallel with the foliation. South-east of N. Grönhult (623/316) aggregates of mainly biotite and quartz megacrysts have given the rock a spotty appearance. Here, larger biotite, muscovite and chlorite flakes are concentrated into aggregates near the quartz megacrysts, which are often polygonized. The megacrysts and the aggregates are probably products of metamorphic differentiation.



Fig. 1. Phenocryst-rich, intensely metamorphosed, weakly foliated (664/868). Intensely metamorphosed, weakly foliated plagioclase (white) phenocryst-rich metamorphosed, 664/868.

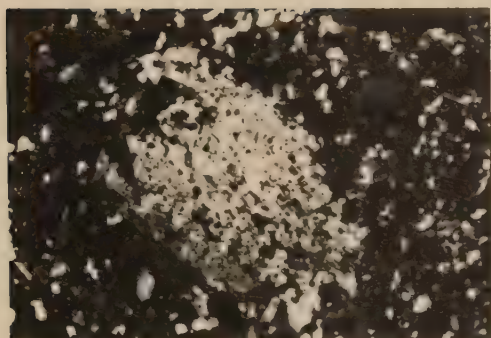


Fig. 2. Plagioclase phenocryst in a metamorphosed metamorphosed. The phenocryst is partly replaced and marginally granulated. 3 mic., 10x (664/868).

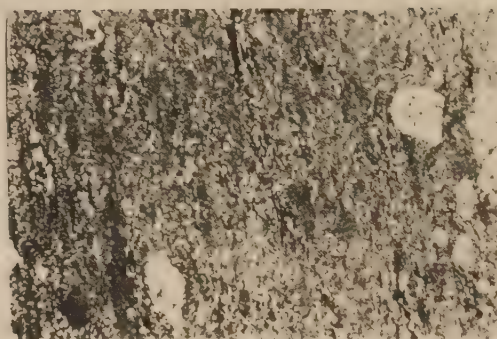


Fig. 3. Acid metamorphosed with phenocrysts (white) of quartz, quartz (white) phenocrysts in a acid metamorphosed (this section coloured). 1 mic., 4x (664/244).

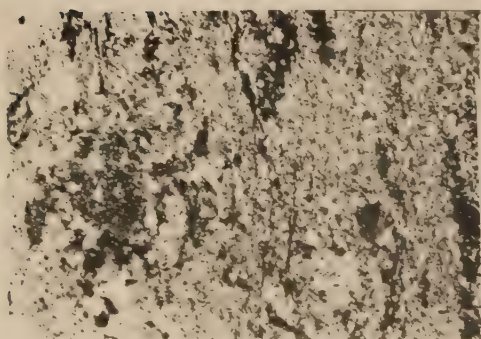


Fig. 4. Plagioclase phenocryst, mainly replaced (middle) and enveloped by metamorphosed. 1 mic., 4x (664/868).

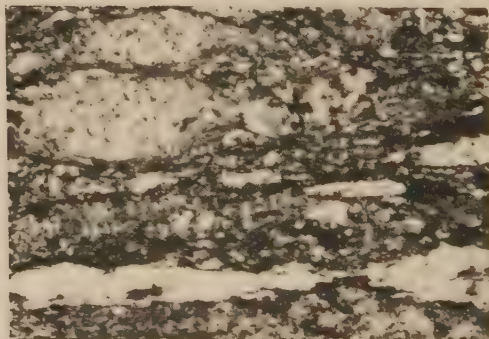


Fig. 5. Metamorphosed, strongly foliated, highly textured with lenses of quartz (white) and feldspar (grey). 1 mic., 4x (664/868).

In other regions, the phenocrysts are replaced by pre-tectonic coarser-grained aggregates, which apparently represent totally recrystallized phenocrysts or phenocryst accumulations. When the acid metavolcanics are more intensely recrystallized, as for instance at locality 641/343, the grain-size becomes coarser and the texture non-porphyritic, sub-polygonal.

In some areas the acid metavolcanics are strongly modified by deformation. The foliation is emphasized by orientated fibrous mica, which is often stressed into thin streaks. Streaks of mica envelop the feldspar phenocrysts, which are strongly sericitized and elongated parallel with the foliation (Fig. 5). Pressure shadows in connection with the phenocrysts are not uncommon.

Tuffites and Metasediments

Tuffites form a transitional stage between metavolcanics and metasediments. Macroscopically they are grey to light grey or yellowish grey.

A thin and repeated banding is parallel with a well developed foliation. Yellowish grey or light grey bands have higher contents of muscovite, while darker, grey bands more or less corresponds to the normal metavolcanics.

As in the case with the strongly foliated metavolcanics, these rock-types have higher contents of mica. The texture is lepidoblastic, but a few secondary flakes of muscovite are orientated randomly. Polygonized, lens-shaped quartz phenocrysts and feldspar aggregates are enveloped by mica flakes and are obviously pre-tectonic. Parallel to the foliation there is sometimes a weak cleavage, which creates rough sericite-coated surfaces.

In the light-coloured bands, muscovite appears as a main mineral (>40 %). Essential minerals are quartz and biotite. Epidote, chlorite, sphene, apatite, sericite and feldspar are subordinate minerals or accessories. The biotite is partly and sometimes completely altered to chlorite.

The parallel arrangement of muscovite and biotite and their intergrowth may indicate a simultaneous and syntectonic crystalli-

zation. Superimposed on this parallel orientation of mica is a later micro- or minor-folding (crenulation foliation). The micas, which define the micro-folds are slightly bent, show undulous extinction, or are broken. More intense deformation and folding causes further rotation of the micas into the axial plane direction S_3 . Onto the micro-folding there is thus superimposed a strain-slip-cleavage, which is parallel or subparallel to the axial-plane of the microfolding. Secondary quartz veins parallel with the foliation S_1 are due to the strain-slip-cleavage micro-faulted or show signs of boudinge.

Thin, white, very light grey or pale red layers of metasediments occur within the metavolcanics. The abundance of the sedimentary layers seems to increase towards the west. They are most frequent close to the metabasite (see map). The maximum thickness of the layers has been estimated to be around 25 m. In spite of their relatively small thicknesses it is difficult to distinguish and trace the individual layers in a lateral direction. The mica contents may vary, but when present in sufficient amounts, the mica gives the rock a foliated appearance. The rock composition varies between that of mica-schists and arkosic sediments. The transition between the sediments and the metavolcanics may be gradual or distinct.

The metasediments are normally white to very light grey, fine-grained mica-schists. The main minerals are quartz and muscovite. Subordinate or accessory minerals are alkali feldspar (partly perthitic), plagioclase, epidote, greenish biotite, and iron oxides. The fabric is a granoblastic to lepidoblastic arrangement of subpolygonal to slightly sutured grains with an average grain-size of 0.1 mm. Coarser-grained quartz is undulatory and is usually marginally granulated. A weak and infrequent crenulation cleavage has induced a diffuse strain-slip cleavage.

A different type of metasediment crops out east of Viken (643/364). That rock is pale red, fine- to medium-grained. The main minerals are quartz and alkali feldspar. Subordinate minerals or accessories are plagioclase, muscovite, biotite, zoisite, epidote, opaques, and garnet.

The texture is somewhat cataclastic. The rock often contains fractured or polygonized megacrysts of quartz and alkali feldspar. Deformation twins, cross-hatched microcline, and perthites are common. The biotite is strongly altered and due to secondary alteration products often semi-opaque.

The contacts with the country rock are sharp, diffuse or gradual and are always concordant with the major lithological units of the area. Stratigraphically this layer belongs to the basal part of the metavolcanic formation, where interlayered accumulations of epiclastic origin, such as mica-schists and micaceous quartzites, are rather frequent. This, and the circumstance that the feldspar/quartz ratio is close to or above 1, which fits very well with arkosic sediments, make it reasonable to assume an epiclastic origin of the layer. Moreover, if the influx of epiclastic material came from a plutonic or high-metamorphic terrain and was of constant rate, the arkosic rock-type represents a much more rapid accumulation than the quartz-rich sediments. These alternations of almost contemporaneous volcanism and sedimentation are in good agreement with the dualism of volcanic and tectonic activities in eugeosynclinal regions.

Mylonite

In strongly tectonized parts of the metavolcanics thin "layers" of mylonite have been found. At UTM 633/343 mylonite crops out as a "layer" with a maximum thickness of 4 m. Table 2, p. 16, shows the mineral composition of the matrix. The mylonite is yellowish grey with a very weak cleavage. In the very fine-grained matrix, here are pink, irregular alkali feldspar-rich and somewhat elongated (1-2 dm in length) fragments.

Microscopically the foliation is to some extent marked by subparallel sericite flakes. The texture is subgranoblastic-polygonal. Quartz and sericite are the main minerals. Accessories are apatite, zircon, and leucoxene. The quartz is sometimes diffusely undulatory and the sericite is occasionally accumulated into aggregates. The somewhat coarser-grained alkali feldspar is most often perthitic.

Intermediate metavolcanics

Intermediate, phenocryst-bearing metavolcanics are stratigraphically superimposed on the acid metavolcanics. Although there are no sharp boundaries between the acid and intermediate metavolcanics, microscopic and chemical analyses have made it possible to distinguish between the two units. Usually the intermediate metavolcanics are somewhat darker and have a higher content of darker minerals. This rock-type may be traced as a relatively thin bed from SE of N. Grönhult to NW of Skärsnäs (see map). Quantitatively they are less important than the acid metavolcanics.

The intermediate metavolcanics are medium dark grey to medium light grey. Phenocrysts are invariably present, but not always macroscopically visible. They never exceed 30 % of the rock volume (Fig. 1). The foliation is weak or absent and a diffuse lineation can be observed locally. Parallel to the foliation there may appear secondary, thin (1-2 cm) quartz-feldspar-rich light bands or lensoids with obscure contacts to the bedrock. The femic minerals are often assembled into aggregates, which give the rock a spotty appearance.

Typical intermediate metavolcanics are represented by analyses 5 and 6 in Tables 1 and 2. The main minerals are quartz and plagioclase. Alkali feldspar, biotite and epidote are essential minerals. Subordinate or accessory are muscovite, chlorite, common hornblende, opaques, apatite, zircon, and sphene (leucoxene).

The phenocrysts are sericitized and often poikilitic, carrying inclusions of quartz and epidote. The grain-boundaries are irregular and the individual phenocrysts are sometimes polygonized. The groundmass consists mainly of quartz, alkali feldspar and plagioclase. The alkali feldspar is usually perthitic. All feldspars have straight to curved grain boundaries and form a granoblastic-polygonal texture. Coarser are more or less lens-shaped feldspar aggregates with characteristic mosaic textures.

The biotite is normally green, but is sometimes bleached and chloritized. Inclusions of leucoxene or epidote are common. Although strongly poikilitic the amphibole is often idiomorphic

Table 1. Modal analyses (volume percent) of acid and intermediate metavolcanics from the Västana area.

Analysis No.	2	3	4	7	5
UTM co-ordinates	632/328	633/354	664/244	625/348	623/351
Quartz	24	26	30	24	23
Plagioclase ¹⁾	36	30	34	37	38
K-feldspar ²⁾	28	23	24	12	13
Muscovite ³⁾	3	5	3	4	+
Biotite	7	11	8	11	14
Chlorite	-	-	-	+	+
Hornblende	-	-	-	-	+
Epidote	+	3	1	8	8
Calcite	-	-	-	+	-
Opaque	1	+	+	2	2
Apatite	+	+	+	+	1
Zircon	+	+	+	+	-
Sphene ⁴⁾	+	1	+	2	+
An mol. p.c. in plagioclase	3-15	12-16	24-26	18-22	20
An mol. p.c. in phenocrysts			32-34		
Average grain size (mm)					
Matrix	0.07	0.08	0.1	0.09	0.04
Phenocrysts		1.5	0.8	1.4	1.0

+ Indicates accessories (i.e. less than 0.5 % of the mode)

1) Including secondary epidote + sericite

2) " perthitic K-feldspar

3) " sericite

4) " leucoxene

No. 2. Acid metavolcanic, medium light grey, without phenocrysts.

" 3. " " , medium grey, with sparse phenocrysts.

" 4. " " , medium grey, with granulated phenocrysts.

" 7. Intermediate metavolcanic, medium dark grey, with phenocrysts.

" 5 " " , medium dark grey, with granulated phenocrysts.

Table 2. Modal analyses (volume percent) of intermediate metavolcanics, "tuffites", and interbedded meta-sediments from the Västana area.

Analysis No.	8	6	9	10	11
UTM co-ordinates	621/355	625/353	629/345	627/355	633/343
Quartz	21	25	23	31	49
Plagioclase ¹⁾	44	37	35	19	-
K-feldspar ²⁾	12	11	12	16	8
Muscovite ³⁾	3	2	1	28	32
Biotite	13	13	27	+	-
Chlorite	-	+	+	+	-
Epidote	1	9	1	5	11
Hornblende	-	-	-	-	-
Calcite	-	-	-	-	-
Opaques	4	1	+	+	-
Apatite	1	+	+	+	+
Zircon	+	+	+	+	+
Sphene ⁴⁾	1	1	+	+	+
An mol. p.c.					
in plagioclase	18-24	20-22	5-22	3-18	
An mol. p.c.					
in phenocrysts	28-38		24-26		
Average grain size (mm)					
Matrix	0.12	0.07	0.09	0.10	0.08
Phenocrysts	2		1.6		

+ Indicates accessories (i.e. less than 0.5 % of the mode).

1. Including secondary epidote + sericite.

2. " perthitic K-feldspar.

3. " sericite.

4. " leucoxene.

No. 8. Intermediate metavolcanic, medium grey, with phenocrysts.

" 6. " " , medium light grey, with strongly granulated phenocrysts.

" 9. " " , medium dark grey, with phenocrysts.

" 10. "Tuffite", mylonite, yellowish grey, with quartz phenocrysts.

" 11. Micaceous quartzite, schistose, yellowish grey, with "fragments."

Table 3. Chemical analyses (w.p.c.) and Niggli values of some acid and intermediate metavolcanics from the Västana area.

Analysis No.	1	2	3	4	5	6
UTM co-ordinates	631/347	632/328	633/354	664/244	623/351	625/353
SiO ₂	70.8	70.4	67.3	66.9	65.5	64.0
TiO ₂	0.47	0.46	0.49	0.52	0.79	0.77
Al ₂ O ₃	14.9	15.5	15.4	16.2	15.6	15.6
Fe ₂ O ₃	1.57	1.93	1.66	2.10	3.80	2.54
FeO	1.31	0.70	1.76	1.50	1.32	2.40
MnO	0.06	0.05	0.07	0.08	0.13	0.11
MgO	0.73	0.57	1.03	1.10	1.67	1.66
CaO	2.00	3.15	2.38	2.90	3.27	5.43
Na ₂ O	2.73	4.06	3.48	3.50	4.43	2.64
K ₂ O	5.44	3.28	4.74	3.80	3.76	4.25
LOI	0.60	0.35	0.84	0.80	0.47	0.95
BaO	n.d.	0.13	0.11	0.13	0.11	0.08
Sum	100.6	100.6	99.3	99.5	100.9	100.4
si	346	327	300	290	252	239
qz	+27	+115	+86	+89	+49	+60
al	43	42	41	41	35	34
fm	17	14	19	20	25	24
c	10	16	12	14	14	22
alk	30	28	29	25	26	20
mg	0.31	0.28	0.35	0.36	0.37	0.38
k	0.56	0.34	0.47	0.41	0.35	0.51
ti	1.7	1.6	1.6	1.7	2.3	2.2
h	10	5	13	12	6	12
w	0.51	0.71	0.45	0.55	0.72	0.48
T	+13	+14	+12	+16	+10	+15
t	+3	-1	0	+2	-4	-7

No. 1. Acid metavolcanic, medium grey, with phenocrysts.

and connected with secondary quartz-feldspar aggregates. Chemical and modal analyses are given in Tables 1 to 3. According to Rittman (1952 and 1962) the metavolcanics are rhyolites or quartz-latites of a Pacific (calc-alkaline) character. The weak alumina excess in analyses 1 and 4 indicates a deviation from the assumed original composition, which is probably due to contamination by argillitic material.

It is evident that potassium in general predominates over sodium and that the quartz and plagioclase percentages are rather constant. From analyses 4 to 7 the alkali feldspar decreases from 24 to 12 volume percent. This variation in mineral composition is not reflected in the chemical classification of the metavolcanics according to Rittman (1952). Using the modal analyses as a basis for the classification of the metavolcanics, different results are obtained, which illustrate the modal alkali feldspar variation in a better way. The chemical classification according to Rittman is useful, although it appears that the ranges of quartz-latites and rhyodacites in his diagrams extend too far towards the basic extreme of series. Nockolds' (1954) and Streckeisen's (1967) classification schemes show good agreement. Applying Streckeisen's diagram to the acid and intermediate metavolcanics (Fig. 6, p. 19) these are simply classified as rhyodacites and dacite.

In structural regard the metavolcanics with interbedded layers of metasediments have a rather regular appearance. The NW - SE foliation is fairly well marked and the dip is normally steep to vertical. Bands reflecting variations in mineral composition are parallel with the foliation. If this compositional layering is a primary feature, S_0 and S_1 are parallel. Evidence of metasomatism leading to compositional layering has not been observed.

Among the planar structures, S_1 is the most conspicuous feature in almost all outcrops. The biotite is obviously syntectonic in regard to S_1 . A later deformation has been traced in the mica-rich interlayered metasediments as a diffuse strain-slip cleavage, which is obliquely orientated to S_1 .

The metavolcanic unit is completely recrystallized. Apart from phenocrysts and small feldspar aggregates, which may be remnants of

Table 4. In order to compare the volcanics' nomenclature of Rittman (1952) and Streckeisen (1967), this table classifies the analyses in Table 1 and 2. To show the difference between the modal and chemical classification of volcanics (Rittman 1952), the data of Table 3 have been added.

Analysis No.	Modal analyses		Chem. anal.
	Streckeisen	Rittman	Rittman
2	rhyodacite	quartz-latite	quartz-latite
3	"	"	rhyolite
4	"	"	quartz-latite
5	dacite	dark rhyodacite	"
6	"	" "	"
7	"	" "	-
8	"	rhyodacite	-
9	"	dark rhyodacite	-

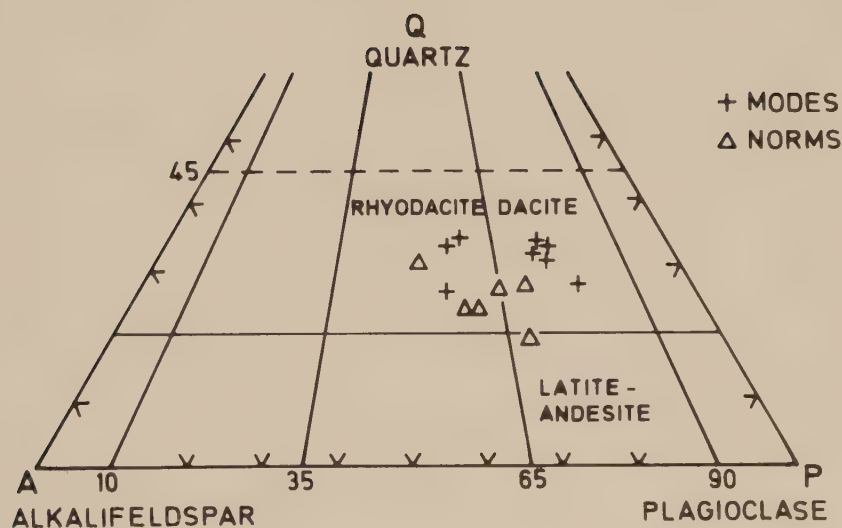


Fig. 6. Modal and normative plagioclase-alkali feldspar-quartz compositions of some metavolcanics from the Văstana area. Nomenclature according to Streckeisen (1967).

fragments, primary features are absent. Due to the sparse outcrops the thickness or total amount of the metavolcanics in relation to the tuffites and metasediments is difficult to estimate. Where best represented the whole sequence seems to be at least 4 km thick. The considerable extension and the relatively high frequency of tuffites and metasediments make an interpretation of the metavolcanics as lavas unlikely. Nevertheless, due to the sparse outcropping, the presence of small lava flows cannot be excluded. The intimate connection between metavolcanics and metasediments indicates a close relationship in time. As already mentioned, the metasediments are more frequent at stratigraphically higher levels, thus indicating a change of facies with decreasing volcanic activity.

Metabasites

Partly amphibolitic metabasites appear in several parts of the mapped area. The largest occurrence extends from the region of Blistorpasjön to Skärsnäs. The lateral extension is there more than 6 km, the maximal width about 100 m.

Relatively thin (1-10 dm) intercalations of metabasites, most often concordant, appear in the metavolcanic formation. On the basis of mineral composition some of them can be called amphibolites, but normally they have the same mineral composition as the large layer. These thin, rather homogeneous metabasites are dark grey to greenish grey, fine-grained and often contain epidote and/or quartz in a network of fine veins or as small aggregates. Foliation is weak or absent.

Although the majority of the thin metabasites is concordant and thus should be regarded as sills or supracrustals, some of them have undoubtedly the character of dykes. Macroscopically, the more or less continuous large layer (see map) consists of fine-grained to very fine-grained, dark grey to dark greenish grey, normally weakly foliated metabasite. However, deviations from the above described normal appearance are common.

Banding may appear and is parallel with the foliation. Due to the very diffuse nature of the banding, it is usually impossible to decide whether it reflects a primary feature or is the result of metamorphic segregation. However, at UTM 634/311 the banding is due to a rather regular alternation between a fairly unaltered metabasite and quartz-feldspar-rich stripes, which constitute about 10 % of the total rock volume. This type of banding is probably a result of tectonic and metamorphic processes and not an original feature.

Another conspicuous trait is the presence of irregular, flattened lenses, which are very much like fragmental material (Figs. 9 and 10). They vary in size between 1 and 20 cm in length and 1 to 20 mm in width. The contacts with the groundmass show all gradations from distinct to diffuse. The difference in composition between "fragments" and matrix is small.

Other inhomogeneties are assemblages of feldspar, quartz, and magnetite. These aggregates may sometimes contain small amounts of garnet. In some outcrops are epidote minerals concentrated into rounded or elongated pods, which have later been cut off by fracture-veins healed by quartz.

The metabasites are chiefly composed of plagioclase, biotite, and amphibole with minor quantities of muscovite, epidote, and magnetite (Table 5, p.25). Texturally the metabasites are characterized by a granoblastic - subpolygonal matrix with coarser-grained biotite and amphibole. When present, the foliation is always marked by parallel biotite flakes, whereas the amphibole may show either a nematoblastic or a decussate appearance. The amphibolites are texturally very much like the metabasites, but the feldspar minerals are not orientated.

The amygdules of the metabasites (Fig. 7, p.22) are usually less than 15 mm in diameter and ellipsoidal in shape. Some of them are zoned and contain feldspar, magnetite, and amphibole (the mineral in the core mentioned first). They are mantled by bleached rims (Fig. 8). Other constituents in segregations or fissures are quartz, alkali feldspar (perthitic), epidote, biotite, muscovite, and sphene. In the matrix the plagioclase is xenoblastic or hypidioblastic and equidimensional. The anorthite contents range between 25 and 45 %.

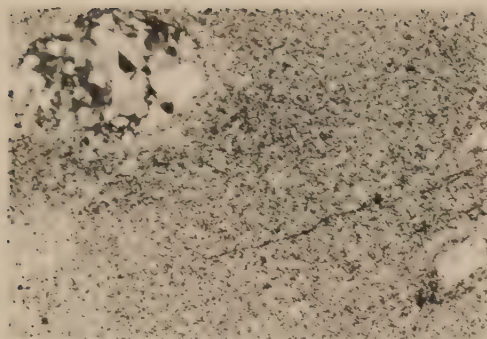


Fig. 7. Metabasite with lensoid shaped amygdaloids. 2 mic., 2x (603/449).

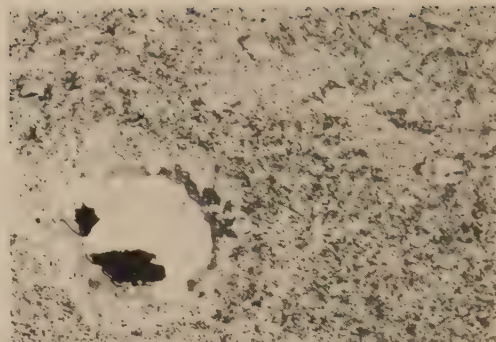


Fig. 8. Metabasite with a zoned amygdale. Feldspar and quartz in centre, surrounded by Biotite (black) and amphibole. 1 mic., 2x (603/449).



Fig. 9. Metabasite with fragmental enclosures (lower part, right). (635/311)

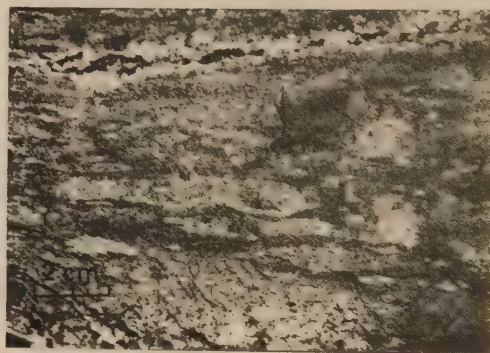


Fig. 10. Metabasite with segregations (upper part) and fragmental enclosures (middle part). (635/311).

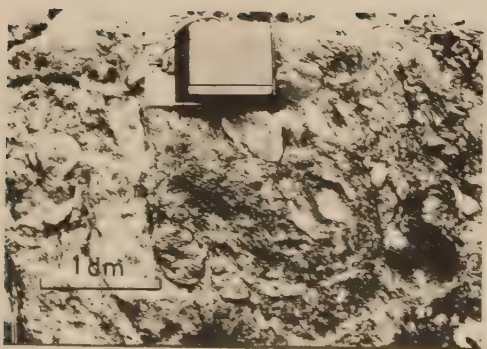


Fig. 11. Metaformational conglomerate with pebbles of quartzite (603/449).

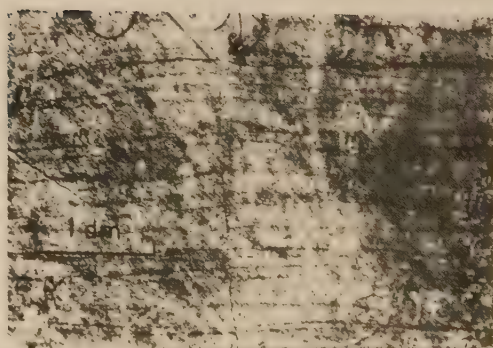


Fig. 12. Laminar heavy minerals in the basal quartzite. (637/306).

Near the contact to the granitoid the plagioclase is more or less sericitized with concomitant albitization. Here, much of the ground-mass consists of microcline (mostly perthitic) and the porphyroblasts are either perthite or antiperthite.

The amphibole is typically an ordinary green to yellowish green hornblende, but bluish green varieties also occur. Close to the amygdules or secondary segregations, the hornblende becomes coarser-grained and poikilitic, often containing inclusions of magnetite. In some cases the amphiboles are partly interlaced with biotite and some aggregates seem to have been completely replaced by fibrous masses of chlorite.

Fissure fillings are in several cases cut by long hornblende prisms piercing in from the matrix. These prisms are sometimes well aligned with the foliation. In contrast, other well-developed crystals may be dispersed at random or aggregated into groups, which often form "rosettes" traversing the parallel orientation. Thus the hornblende may have crystallized during a very late stage of metamorphism when the movements had considerably weakened or perhaps terminated altogether.

The biotite is green to pale green in the metabasites and pale brown in the amphibolites. In the amphibolites the biotite is partly altered to chlorite. Leucoxene occurs along the cleavage planes and there are inclusions of sagenite.

Epidote is a subordinate mineral showing weak pleochroism. The crystals are anhedral, often cloudy, and have cores of allanite. Optical data show that the epidote mostly has the composition of pistacite. Accessories are chlorite, apatite, zoisite, sphene (leucoxene), zircon, allanite and occasionally calcite, microcline, and garnet.

The variation in the major-element composition of the metabasites is described in Table 6, p. 26.

Presuming an igneous origin of the metabasites they are (acc. to Rittman 1952) varying between olivine-andesine-trachybasalt and latite, which should be compared with Streckeisen's terminology (Fig. 13, p. 28). The analyses 14, 15 and 16 have, when compared with the other analyses in Table 7, higher c and w values. This may indicate

a difference of origin, although there is a close association in time. The cross-cutting amphibolites, which have an intrusive character, indicate the presence of at least two generations of basic rocks.

If the metabasites are of a sedimentary origin, they must certainly consist of mixture of pelites and limestone or dolomite, or both. These mixtures, when metamorphosed later may have produced metabasites of virtually the same composition as basic igneous rocks.

Of the analysed rocks metamorphic segregation and metasomatism have at least affected 667/263 and 634/316 (15 and 17). In the latter there is a late K-metasomatism with replacement of Na by K. Only one of the analyses has a k-value exceeding 0.50, whereas mixtures of clay and dolomite normally have very much higher k-values. In addition, although some pelites have TiO_2 contents of about 2 per cent, they are generally poorer in titanium. Shaw (1956) showed that the average pelite contains 0.8 per cent TiO_2 . Thus, when pelites are diluted with limestones or dolomites, which are very poor in titanium, the resulting TiO_2 contents are obviously bound to be much lower. The high T and low t values indicate that there is no or only minor contamination of argillitic material.

Trace-element analyses of Ni and Cr show no significant trends.

However, successful attempts to distinguish between ortho- and para-amphibolites have been made by Evans and Leake (1964). From Fig. 14, p.28, it is evident that the metabasites are close to the igneous trend of the Karoo Dolerites. Due to the low number of analyses, however, this is not conclusive. However, in conjunction with the above mentioned data this plot can be used as an indication of origin. If sedimentary metabasites are to include mafic tuffs, then clearly no chemical criteria can be expected to reveal differences between the two rock types. The lack of gradation between the metabasites and country rocks and the general absence of carbonate remnants within the metabasites fail to support an origin from mixtures of sediments. Although most of the banding present in the main occurrence is due to metamorphic segregation, the presence of primary bands, which now are more or less concealed by later metamorphic events, cannot be excluded.

Table 5.

Modal analyses (volume percent) of metabasites
from the Västana area.

Analysis No.	13	14	18	19	17
UTM co-ordinates	628/343	660/247	623/349	641/349	634/311
Quartz	-	+	2	+	3
Plagioclase ¹⁾	38	34	44	19	38
K-feldspar ²⁾	-	8	+	-	2
Muscovite ³⁾	1	12	1	10	+
Biotite	34	6	24	20	53
Chlorite	+	+	-	-	-
Epidote	2	4	4	2	+
Common hornblende	17	22	18	46	+
Calcite	+	-	-	-	-
Opaque	6	8	4	+	2
Apatite	+	+	+	+	+
Sphene ⁴⁾	+	5	+	1	1
An mol. p.c. in plagioclase	25-32	35-38	32-36	42-45	38-40
Average grain-size (mm)	0.09	0.15	0.11	0.08	0.10

+ Indicates accessories (i.e. less than 0.5 % of the mode).

1. Including secondary epidote + sericite.

2. " perthitic K-feldspar.

3. " sericite.

4. " leucoxene.

No. 18. Metabasite, medium dark grey, very fine-grained foliated.

" 19. Amphibolite, dark grey, fine-grained.

Table 6. Chemical composition (weight percent) of some metabasites from the Västana area.

Analysis No.	12	13	14	15	16	17
UTM co-ordinates	626/345	628/343	660/247	667/263	621/351	634/311
SiO ₂	48.7	51.4	51.9	51.9	53.5	54.6
TiO ₂	1.98	1.47	1.23	1.38	1.22	1.70
Al ₂ O ₃	16.6	17.6	15.3	16.3	16.0	15.2
Fe ₂ O ₃	3.43	5.36	7.20	6.70	6.08	4.00
FeO	9.06	6.97	2.31	5.90	3.24	7.17
MnO	0.30	0.18	0.16	0.21	0.18	0.15
CaO	4.32	4.87	7.77	6.00	7.12	4.00
MgO	6.44	5.64	4.90	3.90	4.97	3.72
Na ₂ O	3.54	4.30	3.03	5.00	3.91	2.42
K ₂ O	3.89	2.54	3.72	1.00	2.47	4.96
LOI	1.43	0.45	1.02	1.40	0.94	0.68
P ₂ O ₅	0.59	-	0.24	0.70	0.23	0.73
BaO	0.11	-	0.14	0.07	0.10	0.09
Sum	100.4	100.8	98.9	100.5	100.0	99.4

No. 12. Metabasite, dark grey, fine-grained weakly foliated with quartz and epidote veins or aggregates.

" 13. Metabasite, dark grey, fine-grained weakly foliated with vague banding and partly epidotized.

" 14. Metabasite, dark greenish grey, fine-grained in contact with granite.

" 15. Metabasite, medium dark grey, very fine-grained weakly foliated and banded.

" 16. Metabasite, dark greenish grey, fine-grained, partly scoriaceous.

" 17. Metabasite, dark greenish grey, very fine-grained, banded.

Table 7. Niggli values for metabasites from the Västana area

Analyses No.	12	13	14	15	16	17
si	120	129	138	138	143	163
qz	-38	-29	-18	-20	-14	-3
al	24	26	24	26	25	27
fm	50	46	39	43	40	44
c	12	13	22	17	21	13
alk	15	15	14	14	14	16
mg	0.47	0.45	0.49	0.36	0.49	0.37
k	0.41	0.27	0.44	0.11	0.29	0.57
ti	3.7	2.8	2.5	2.7	2.5	3.8
h	12	4	9	12	8	7
w	0.25	0.40	0.73	0.50	0.62	0.33
T	+10	+12	+10	+11	+11	+10
t	-2	-2	-12	-6	-10	-3

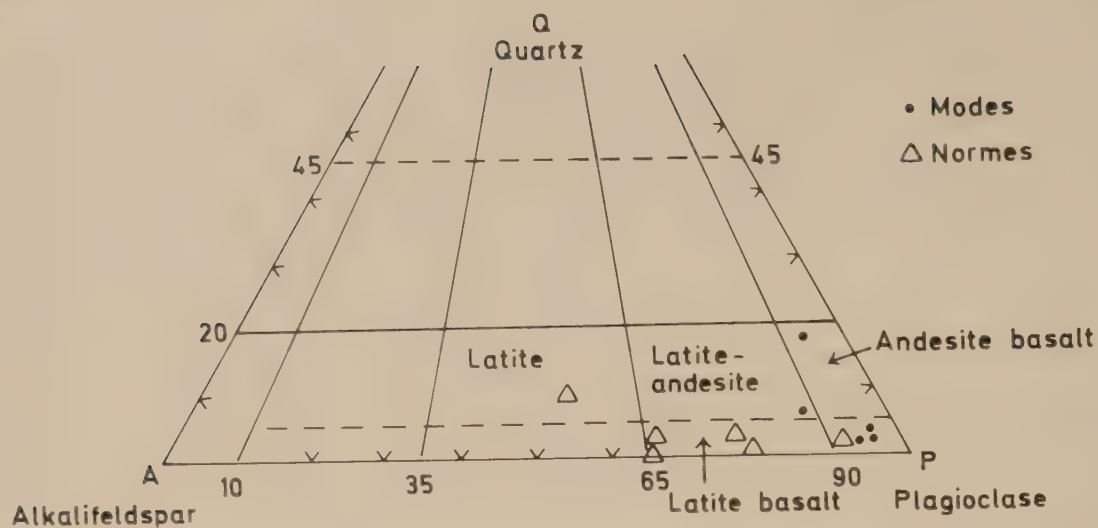


Fig. 13

Classification of metabasites and amphibolites
from the Västana area all to streckeisen (1967)

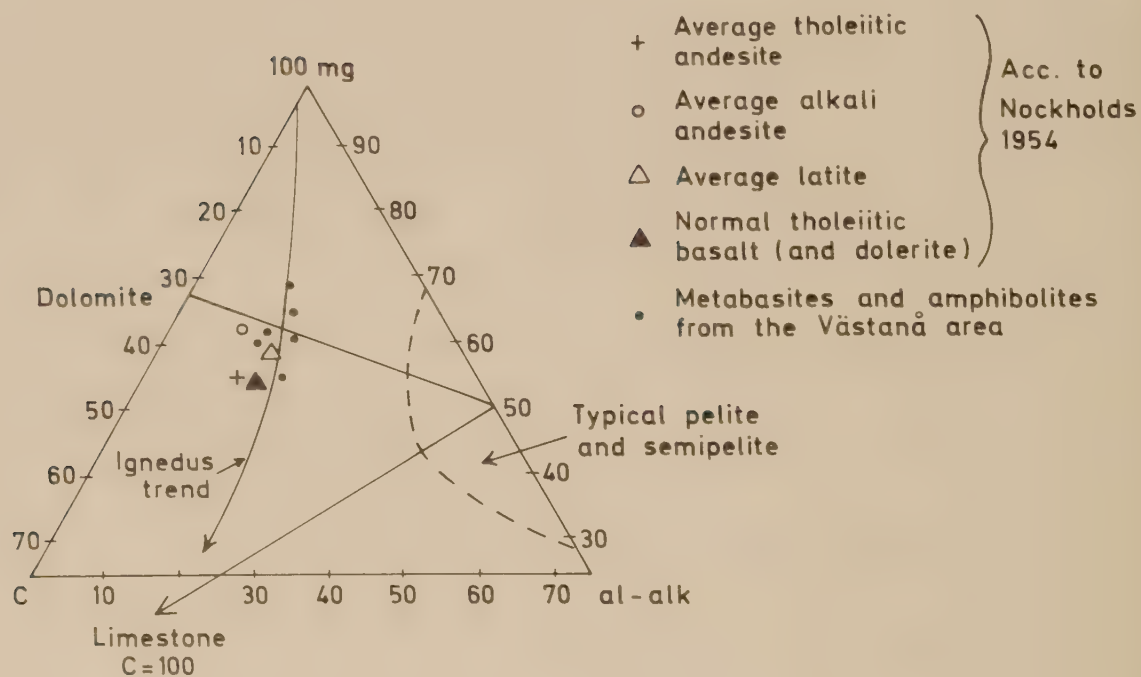


Fig. 14

Part of the 100 mg, C and (al - alk) triangle.
(after Leake 1964)

It is thus suggested that the metabasites are products of a basic volcanic activity of an essentially tuffaceous character. The scoriaceous appearance of the metabasite at 621/351 indicates the presence of lava of spilitic character. Moreover, the interbedded metabasites may also be sills or feeder channels. The cross-cutting amphibolites, like the amphibolites in the quartz-feldspar-gneiss, are clearly intrusive with a dyke character.

Metasedimentary formation

Metasediments occupy the central part of the mapped area. These sediments can be followed from N of Klagstorp to WNW of Skärsnäs and have a maximal width of about 2.5 km. Micaceous quartzite and micaschist are the main members of the unit and completely dominate the northern part of the formation. South of the great fault (see map), the basal member of the formation is a micaceous quartzite, which grades into a mica-schist. These members are upwards followed by an orthoconglomerate - orthoquartzite - micaceous quartzite - mica-schist - intraformational orthoconglomerate and then alternating mica-schists and micaceous quartzites, which also, northwest of Lake Bären, include a thin layer of reddish meta-arenite.

The transitions between the members are often gradual. Within the members, compositional layering is not unusual.

Orthoquartzite

Orthoquartzite (cf., Krynine 1948) occurs in the metasediments predominately south of N Grönhult. The lowermost quartzite may be followed from the southernmost part of the metasediment formation to just south of N Grönhult and has a maximal width of 100 m. The unit is superimposed upon micaceous quartzite, which seems to decrease in thickness northwards. In the quartzite, nearly all outcrops contain sometimes closely spaced, thin layers of hematite, sphene, and tourmaline (Fig. 12, p. 22). These layers or laminae represent bedding-planes, which are often cross-bedded.

Thin beds of quartzite have also been found in the higher stratigraphic levels of the formation. The thickness and frequency of the beds seem to decrease towards the higher stratigraphic levels where they can only be traced as thin intercalations of a width of 1 - 2 cm.

Very light grey to light greenish grey colours predominate. The texture is varying from xenoblastic to granoblastic-polygonal and the grain-size varies from 0.07 - 2 mm with an average size of 0.14 mm. Coarser grains have most often irregular or sutured grain boundaries and the individuals show undulous extinction or even deformation lamellae crossing the whole crystal. This may indicate a mild cataclasis after the crystalloblastic fabric had formed.

Sericite occurs as an essential mineral (Table 8, p. 34) in sub-parallel, disseminated flakes or in sparse aggregates. Accessories are black tourmaline, sphene, opaques, zircon, and apatite, which are mostly concentrated into thin lamellae tracing the bedding-planes. The tourmaline and zircon have characteristics of an allothigenous origin. Lazulite is occasionally found as irregular aggregates or as fissure-fillings.

Orthoconglomerate

Well-rounded, sometimes slightly deformed quartz pebbles are found in thin layers in the basal part of the formation. The size of the pebbles varies between 1 and 5 cm. Although somewhat elongated, they are not always parallel to the regional foliation. With a few exceptions this conglomerate is monomict.

An intraformational conglomerate (Fig. 11) is found as a thin layer in a micaceous quartzite, which gradually passes over into a mica-schist. This conglomerate is clearly affected by deformation and shows an excellent lineation. The average index of elongation is 1:1.5 with a maximum value of 1:4. The size of the pebbles varies between 1 and 20 cm. Like the basal conglomerate the intraformational one is almost monomict carrying pebbles of quartzite. The non-quartzitic pebbles are dark grey or black and very fine-

grained, and according to Bäckström (1897) represent remnants of tourmaline schist.

These types of conglomerate, which are composed of pebbles of one single rock-type may only represent a minor break in the sedimentation record and are probably the product of a transgression over a lowland plan.

Micaceous quartzite

White, pale red-purple or pale pink micaceous quartzite appears as beds in different stratigraphic positions within the metasediments. The thickness of the units varies from 1 - 2 cm to more than 10 m. Except for thin intercalated layers of quartzite and mica-schists, primary structures seem to be lacking in this member. Macroscopically the metasediments are fine-grained. The foliation is well marked by orientated muscovite flakes, which are parallel with the above-mentioned intercalated layers.

Variation in colour depends on the presence of a pink muscovite and/or a reddish andalusite. Mineral compositions are given in Table 8. The dualism between the muscovite and the aluminium silicates is well stressed. High percentages of mica exclude the presence of andalusite and kyanite.

The texture is subgranoblastic - polygonal. Because of strong recrystallization remnants of clastic grains have not been found. Coarser-grained quartz frequently contains inclusions of sericite and is often undulous with embayed or sutured grain boundaries. The optical data for white muscovite are $2V = 40^\circ$, birefringence 0.041 and for pink muscovite 44° and 0.035 respectively. X-ray investigations of the pink and white types show no difference. Both mica types have a $2M_1$ structure. Coarser poikilitic muscovite flakes are often kinked or bent.

A pink or reddish andalusite can be found in several outcrops in the northern part of the sediments (see map). Locality 618/350, S of Skärsnäs has an especially high content (18 vol.p.) of andalusite. A green variety of andalusite has been found in a boulder N of Lyckås (637/307). This mineral is distinctly pleochroic with

X = emerald green, Y = yellow green and Z = yellow. The optical axial angle is $2V = 70^\circ$. The MnO content is 3.09 w.p.

Bluish kyanite seems to be more frequent than andalusite and can N of Grönhult be traced in several outcrops. The mineral is often concentrated into 1-2 cm large aggregates. Small acicular crystals are often assembled into thin lamellae, which are sub-parallel to the foliation. Due to later structural events, some crystals are deformed or even broken (Fig. 17) ~~2-3~~ .

Subordinate minerals or accessories are sphene (leucoxene), hematite, zircon, magnetite, apatite, and rutile. Leucoxene appears as dark brown almost opaque rims surrounding the opaque minerals, which may be assembled into thin horizons concordant with the foliation.

Mica-schist

This type of rock appears to increase in frequency, towards the west and north. Mica-schist is most common NW of Lake Bären. The colour is usually white or very light grey. The occurrence of sparse aggregates of kyanite and/or chloritoid sometimes gives the schists a spotty appearance. Due to the high mica contents (Table 8), the lepidoblastic texture is well marked. Apart from chloritoid and pyrite, the mineral assemblage is the same as that in the micaceous quartzite. The contents of Al-silicates are much lower than in the micaceous quartzite. Only in a few places NW of Lake Bären do kyanite and andalusite occur in macroscopically detectable amounts.

The main minerals are quartz and muscovite. Subordinate or accessory are kyanite, chloritoid, hematite, zircon, apatite, and pyrite. The chloritoid forms acicular or prismatic crystals often assembled into aggregates. Small grains are sometimes found in thin horizons or stringers parallel with the foliation. Larger individuals have no relation to the foliation. The birefringence has an average of 0.16, $2V$ is 55° to 62° . The pleochroism is well marked: X = pale green, Y = greenish blue and Z = colourless to

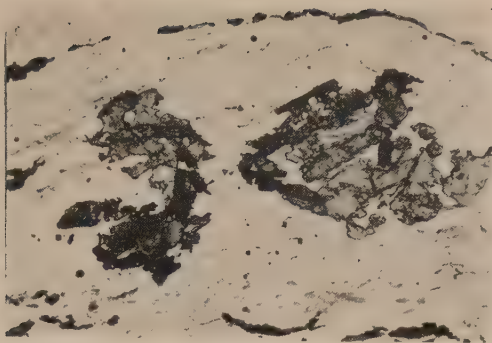


Fig. 15. Larger, random orientated kyanite crystals (middle), which are swept around by thin horizons of kyanite. 1 nic., 16x (612/331).

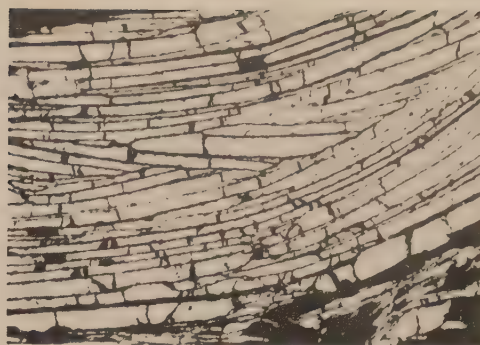


Fig. 16. Specularite (hematite) from the iron ore. 2 nic., 260x (667/281).

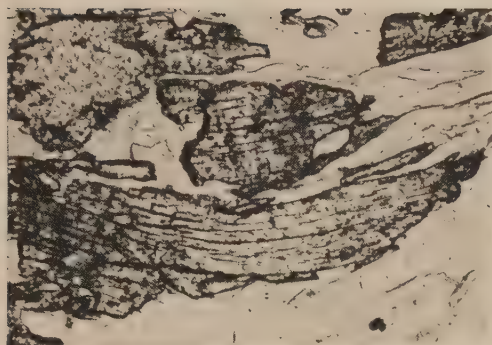


Fig. 17. Deformed crystal of kyanite. 1 nic., 40x (611/331).



Fig. 18. Kyanite with enclosed thin horizons of hematite. 1 nic., 40x (609/335).

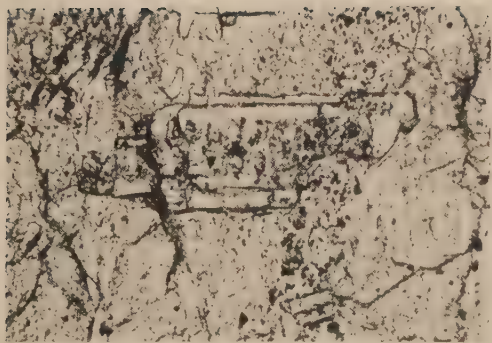


Fig. 19. Kyanite (centre) enclosed by andalusite. 1 nic., 40x (618/350).

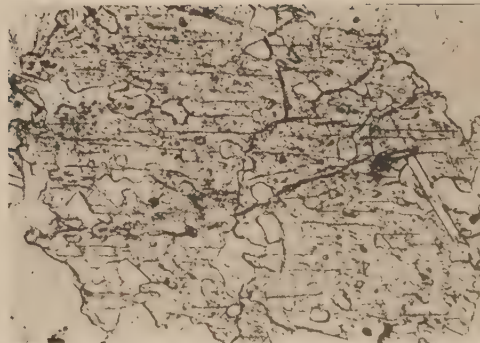


Fig. 20. Poikilitic andalusite. 1 nic., 40x (618/350).

Table 8. Modal analyses (volume percent) of metasediment from the Västana area.

Analysis No.	20	21	22	23	24	25
UTM co-ordinates	618/350	607/339	612/331	611/331	620/327	637/307
Quartz	44	61	79	75	74	91
Muscovite ¹⁾	10	37	13	16	25	8
Andalusite	18	-	-	-	-	-
Kyanite	23	-	7	6	-	-
Chloritoid	-	-	-	1	-	-
Epidote	-	-	-	-	-	+
Opaque	4	1	+	1	+	+
Apatite	+	+	-	-	-	+
Zircon	+	+	+	+	-	+
Sphene ⁴⁾	+	+	+	+	+	+
Tourmaline	-	-	-	-	-	+
Rutile	+	-	+	+	-	-
Grain size						
(mm) matrix	0.1	0.12	0.12	0.1	0.1	0.14

+ Indicate accessories (i.e. less than 0.5 % of the mode).

1. Plus sericite.

4. " leucoxene.

No. 20. Micaceous quartzite, pale red purple, with andalusite and kyanite.

" 21. Mica schist, light grey.

" 22. Micaceous quartzite, white with kyanite.

" 23. " " , very light grey, with kyanite and chloritoid.

" 24. Mica schist, pale pink.

" 25. Quartzite, very light grey.

Table 9. Chemical composition and Niggli values of metasediments from the Västana area.

Analysis No.	20	21	22	23	24	25
UTM co-ordinates	618/350	607/339	612/331	611/331	620/327	637/307
SiO ₂	64.4	79.2	84.5	85.9	88.0	95.8
TiO ₂	0.74	0.32	0.18	0.26	0.13	0.10
Al ₂ O ₃	26.0	12.5	11.5	12.0	7.6	2.1
Fe ₂ O ₃	5.18	1.31	0.90	0.23	0.80	0.50
FeO	0.03	0.22	n.d.	0.78	n.d.	0.10
MnO	0.03	0.01	0.01	0.01	0.05	0.01
MgO	0.02	0.09	n.d.	0.01	n.d.	0.03
CaO	n.d.	n.d.	n.d.	n.d.	n.d.	0.10
Na ₂ O	0.21	0.23	0.05	0.05	0.06	0.10
K ₂ O	2.35	4.18	0.88	0.05	2.58	0.40
LOI	1.66	1.76	0.84	0.96	1.13	0.10
BaO	n.d.	0.06	n.d.	0.03	0.04	0.10
Sum	100.6	99.9	98.9	100.3	100.4	99.4
si	307	638	1047	1072	1286	4260
qz	+174	+484	+917	+968	+1086	+4097
al	73	64	84	88	66	55
fm	19	11	9	11	9	23
c	0	0	0	0	0	6
alk	8	25	8	1	25	16
mg	0.00	0.10	0.00	0.01	0.00	0.08
k	0.88	0.92	0.92	0.39	0.96	0.72
ti	2.6	2.1	1.7	2.4	1.4	3.3
h	26	51	35	40	55	15
w	0.99	0.84	1.00	0.20	1.00	0.81
T	+64	+39	+76	+87	+40	+39
t	+64	+38	+76	+87	+40	+33

pale yellow. The optical character is positive, lamellar twinning is common.

Sedimentary structures

As already mentioned, the thin bedding-planes or laminae (< 1 cm) are marked by heavy minerals, frequent in the basal quartzite. In this member, lamination is most common in the stratigraphically lower parts. NE of Lyckås (637/307) parallel laminae are exposed. The maximum distance between the layers is 5 cm. Within the horizons the heavy minerals are concentrated to the very lowest part and then gradually decrease upwards. This phenomenon is well reflected by a gradual change of rock-colour from black to greyish white. In the stratigraphically higher levels, the parallel laminae are almost absent and are replaced by tangential foresets, inclined lamination or cross-bedding.

The tangential foresets observed in the quartzite can be subdivided into two types: 1. Relatively thin and diffuse straight "branches", which, when assembled in more distinct lamellae, are curved into parallelism with the underlying bottomset. 2. Evenly inclined foresets, 5 cm in thickness or more, with very small angles against the top- and bottomsets. This structure can be regarded as low-angle cross-bedding.

According to Potter and Pettijohn (1963) cross-bedding is defined as a structure that is confined to a single sedimentation unit and consists of internal bedding called foreset bedding, inclined to the principle surface of accumulation. Pettijohn (1962, p. 1471) pointed out "that the term is properly applied only to high-angle cross-laminations. These deviate from the normal essentially horizontal bedding, and forms angles of 10° - 30° or more with the true bedding". As already mentioned, the quartzite contains cross-bedding. NE of Lyckås (636/307) several outcrops exhibit units, which are accentuated by thin horizons of heavy minerals. The separate strata vary in thickness from 1 to 50 cm. Set thickness or the cross-bedded unit is invariably large (> 5 cm). The cross-stratified unit normally consists of a single set of cross-strata, which essentially are litologically uniform.

The lower bounding surface of the set is of a non-erosional character and thus planar.

The described structures indicate that the cross-bedded units are of alpha- or beta-type according to Allen (1963). When using the cross-stratified units as a criterion of top-bottom direction, all the observations indicate higher stratigraphical horizons towards the southwest.

Parallel lamination, cross-bedding and tangential foresets all are sedimentary structures, which can be related to a flow regime. Tangential foresets with flat evenly sloping surfaces and parallel lamination are bedforms belonging to the transition zone between the lower and upper flow regime, or to the lower part of the upper flow regime (Harms and Fahnestock, 1965). Moreover alpha- or beta-cross-stratifications are structures of the transition zone, but the separate quartz pebbles in the basal conglomerate underline the presence of an upper flow regime with a relatively rapid water current. Due to the sparse outcropping and relatively few observations of sedimentary structures, attempts to determine paleocurrent directions were regarded as senseless.

The high quartz content is indicative of a high degree of maturity which may depend on the recycling of older mature sandstones or on a considerable transport and winnowing action before the final accumulation. The described sedimentary structures are in accordance with the ones expected in a delta environment.

Hematite ore and associated minerals

Blomstrand (1866) described the occurrence of a thin hematite ore-body. This ore is situated N of Klagstorp (667/281) in the lower parts of the basal quartzite (see map). The thickness is about 2-3 m. The ore body is conformable and consists of hematite in thin plates (specularite, Fig. 16, p. 33) with disseminated quartz and muscovite. Accessories are kyanite, svanbergite, lazulite, apatite, pyrophyllite, and a number of rare minerals. From a locality 500 m south of the occurrence, G. de Geer (1886) reported the presence of thin hematite layers with subordinate magnetite in a quartzitic outcrop.

At the time of Blomstrand's investigation the mine was abandoned, although sporadic mining has been carried out later. Today the mine pit is water-filled and the dumps overgrown. This renders attempts to find new specimens of the rare minerals rather unpromising. Descriptions of these minerals are therefore based on the investigations carried out by Blomstrand (1866 and 1868). Chemical and physical characteristics are given in Table 10, p. 39.

The phosphates and other rare minerals are mostly aggregated into veinlets or nodules, but may also occur disseminated. The processes forming these minerals must have taken place in separate stages. Thus *svanbergite* and *pyrophyllite* are found as microscopic veinlets in *attacolite*, *lazulite* and in one variety of *apatite*. *Pyrophyllite* is chiefly found as fissure-fillings and *apatite* appears as two varieties, which obviously were not formed at the same time. Excluding *lazulite*, the phosphate minerals seem to be restricted to the surroundings of the ore.

The conformable iron ore shows a rapid lateral narrowing. Sparse occurrences of conglomerate and parallel lamination in the surroundings of the mine suggest a sedimentary environment. The relatively thick accumulation of hematite must have formed in very favorable circumstances.

The presence of rare phosphate minerals, *svanbergite*, *pyrophyllite*, and *kyanite* within the mine area made P. Geijer (1963) suggest that: "It is evident that the phosphates and the *svanbergite* represent material epigenetically introduced into the sediments and their occurrence is so markedly restricted to the chief deposit of hematite as to suggest for the latter a similar origin".

The thin hematite layers and the heavy minerals *zircon*, *apatite* and *tourmaline*, which sometimes show detrital shapes, are undoubtedly of sedimentary origin. The primary features indicate a sedimentation belonging to the transition zone between the upper and lower flow regime or to the lower part of the upper flow regime (see p. 37). Accordingly the oxidation regime was fairly strong during the deposition. In the presence of iron-bearing solutions the iron would be precipitated in a suitable physical and chemical environment forming ferric hydroxides or oxides.

Table 10. Chemical composition and physical data of rare minerals in the iron ore area.

	Augelite	Attacolite	Berlinite	Cirrolite	Lazulite	Näsumlite	Svanbergite	Trolleite	Westanite
SiO ₂	-	-	-	-	-	50.91	-	-	43.44
Al ₂ O ₃	49.15	29.75	40.27	20.54	32.82	27.86	34.95	43.26	51.02
Fe ₂ O ₃	0.89	3.98	-	-	-	1.36	-	2.75	1.30
FeO	-	-	0.26	0.91	7.82	-	0.73	-	-
MnO	0.31	8.02	-	2.24	0.18	0.36	-	-	-
MgO	-	0.33	-	0.21	9.05	-	0.24	-	-
CaO	1.09	13.19	-	29.37	0.84	13.82	16.59	0.97	-
Na ₂ O	-	0.45	-	-	-	-	-	-	-
P ₂ O ₅	35.04	36.06	54.84	41.17	43.83	-	15.17	46.72	-
H ₂ O	12.85	6.90	4.14	5.06	5.72	4.39	12.21	6.23	4.24
SO ₄	-	-	-	-	-	-	15.97	-	-
PbO	-	-	-	0.11	-	-	3.82	-	-
Sum	99.33	98.68	103.51	99.61	100.26	98.70	103.50	103.03	100.00
G	2.77	3.09	2.64	3.08	-	-	3.29	3.10	-
Colour	Colourless	pale	colourless	pale	azure	white	pale	pale	red
	to pale	red	to greyish	yellow	blue		yellow	green	
	red								

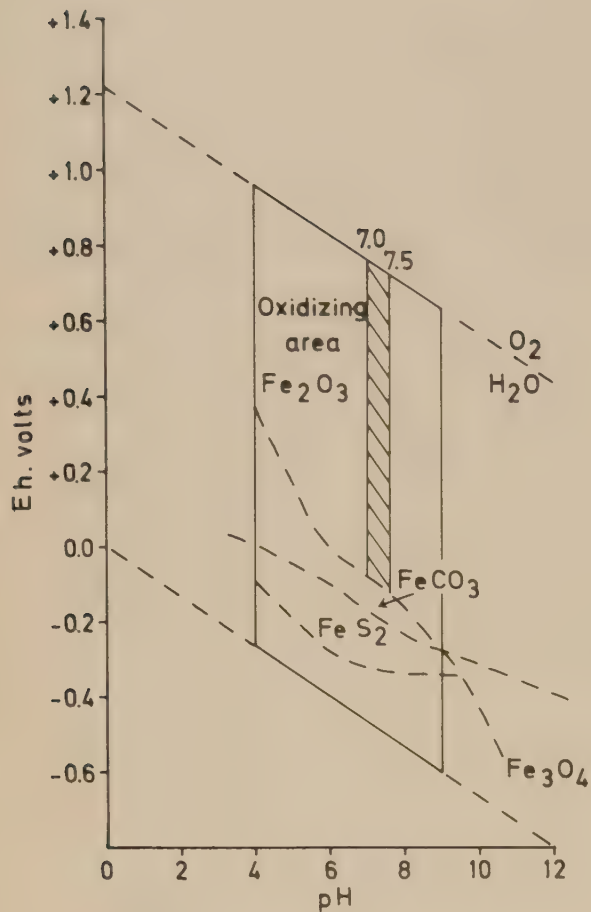


Fig. 21

Eh-pH relations of iron oxides, sulfide, and carbonate, at 25°C and 1 atmosphere.

(From James 1966, p 14)

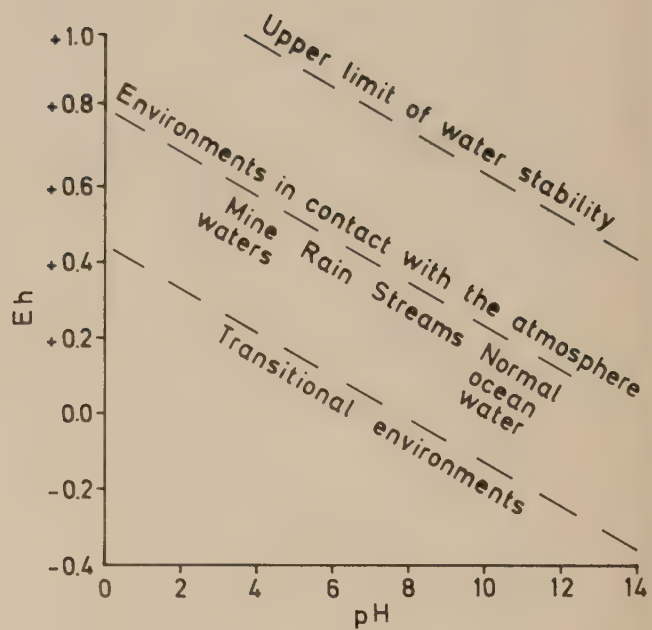


Fig. 22

Approximate position of some natural environments as characterized by Eh and pH.

(From Garrels 1960, p. 201)



Fig. 23

Depositional zones of precipitated iron compounds in a restricted deep basin. (From James 1954, p 242)

The relation between various compounds of iron in solution are mainly controlled by the oxidation potential (Eh) and the hydrogen-ion concentration (pH). In geological processes the range of these factors is restricted, Figs. 21 and 22 showing schematically the probable limits of Eh and pH in near-surface environments. Regarding the mineral assemblage in the mine area, the most conspicuous feature is the presence of rare minerals containing phosphorus, calcium and manganese, but their amounts are small as compared to those of hematite.

The behaviour of manganese in solution is well known and is much like that of iron. Neither P nor Ca is affected by changes in Eh kept within the limits valid in nature, but both elements are strongly pH-dependent. A sediment with a high contents of calcium phosphate can only be formed in a chemical environment where pH values vary between 7.0 and 7.5 (Krumbein and Garrels 1952). The precipitation of sulfates requires fairly oxidizing conditions and will in a pH-Eh-diagram coincide with much of the hematite stability field. Thus a simultaneous precipitation of iron and small amounts of manganese, calcium and sulfates is very possible.

According to the different stability fields of the iron minerals in a pH-Eh-diagram, the sedimentary iron accumulations can be divided into four major groups (Fig. 23). The oxide-facies comprises two sub-facies. One is characterized by hematite and the other by magnetite. Both are primary sedimentary and evidently the hematite facies is formed in strongly oxidizing, near-shore environments.

Consequently, it is not impossible that the final minerals association of the ore are products of metamorphic events, which were essentially isochemical, disregarding possible changes of the H_2O and CO_2 contents.

Gneiss complex

Gneissic, fine-grained, medium grey, pale brown or pale red rocks occupy the north-western part of the mapped area. Outside of this region gneissic rocks have also been found in two outcrops S and SW of Klagstorp. The gneisses are foliated and inhomogeneous on the scale of an outcrop. The compositional variation is expressed

by the abundance of ferromagnesian minerals relative to quartz and feldspar and by the abundance of potash-feldspar relative to plagioclase.

One of the most conspicuous features is the presence of banded and laminated units (Fig. 24, p. 44), which will be further discussed in the following section. Disregarding these structures the change of rock-colour seems to be gradual and has no correlation with the alkali ratios. Although the banding and lamination suggest a supracrustal origin, other "primary" features within the formation seem to be lacking entirely.

In some regions (marked on the map) the gneiss is clearly affected by late tectonic movements induced by faulting. These events have led to a zonal and often strong cataclastic deformation, which will be described farther below.

Under the microscope the foliation is less prominent. Sub-parallel mica flakes appear in a framework of coarser-grained feldspar and quartz, which are strongly granulated when a more penetrative schistosity is developed. In extreme cases the "megacrysts" have been drawn out of different orientation.

The mineral composition (Table 11, p. 46) is simple and the main constituents are quartz, microcline (partly perthitic), and sodic plagioclase. Essential or subordinate minerals are muscovite and biotite. The grain-size varies between 0.04-0.2 mm, ca 0.1 mm being the most common grain-size.

The matrix is somewhat unequigranular with textures varying from subgranoblastic-polygonal to a fabric with distinctly sutured grain boundaries. Both the feldspar and quartz individuals may reach a size of 1.5 mm, but on an average the feldspar is always coarser-grained.

Thin veins and lensoids contain coarser-grain quartz, which shows mosaic texture and is sometimes weakly undulous. Interstitial quartz shows in most of the thin-sections a tendency towards symplectic intergrowth with feldspar.

The feldspar crystals are anhedral with slightly curved or embayed grain boundaries. As already mentioned, microcline is a main constituent, which is often cross-hatch twinned or perthitic.

In specimen 604/335 the plagioclase is partly enveloped by clear rims, which most probably are due to replacement of alkali feldspar. The next step in the transformation of plagioclase can be seen in the patch-perthites with similarly orientated relic blocks of the original feldspar in a K-feldspar host. Twins are common in plagioclase and are often seritized or have inclusions of quartz.

More accentuated effects of dynamic deformation can be found in specimen 605/308, Fig. 27. Here the coarser feldspar crystals are marginally granulated or even disrupted and secondarily healed by quartz. The plagioclase shows wedge-shaped deformation twins and the biotite is almost completely transformed into fibrous, semi-opaque pseudomorphs. In some parts quartz forms elongated, fine-grained, recrystallized aggregates in the matrix, where it is accompanied by feldspar and mica in layers sweeping around large relic "porphyroclasts". This mortar texture, though only partial, indicates along with the low content of matrix, the presence of a protoclastite (Spry 1969).

Thin flakes of muscovite are aggregated into thin streaks, which mostly are parallel with the foliation or are linked with the grain boundaries. Larger disseminated poikilitic flakes are probably of secondary origin.

Green to yellowish green biotite with rugged boundaries is a subordinate mineral in this rock-type. An accentuated orientation is missing. The crystals rather seem to be scattered in between larger feldspar crystals or appear as scaly aggregates. The colour is often uneven and spotty depending on alteration products like opaques, leucoxen and sericite, which normally are concentrated in the grain fringes or the cleavage planes. Pleochroic halos are rare.

Epidote is found as an accessory, anhedral disseminated mineral. The pleochroism is weak, colourless to yellowish green and the fringes are often blurred. Inclusions of allanite are not uncommon. Garnet is a subhedral erratic constituent. Some of the individuals seem to contain alteration products like chlorite and epidote. Other accessories are hematite, secondary Fe-products, sphene (leucoxene), apatite, zircon, and rutile.

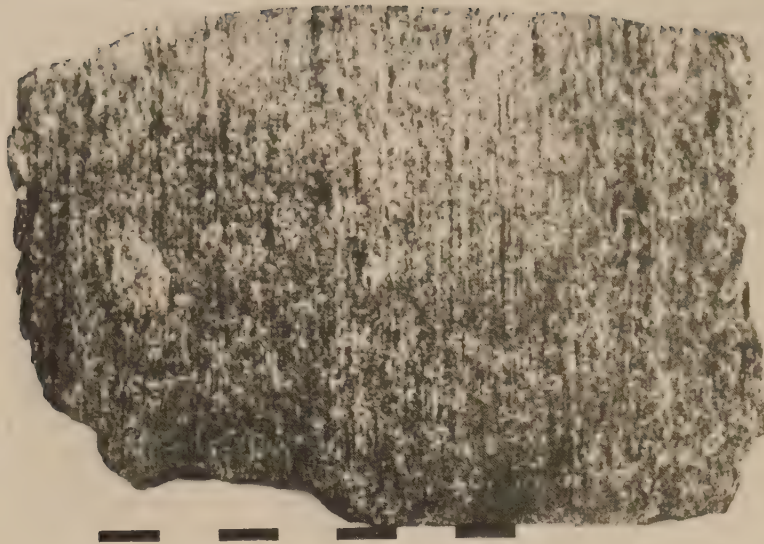


Fig. 24. Non-banded, foliated gneiss. Parallel foliation darker streak of biotite aggregates. (614/319).



Fig. 25. Gneiss with lensoid shaped biotite rich bands. (620/296).

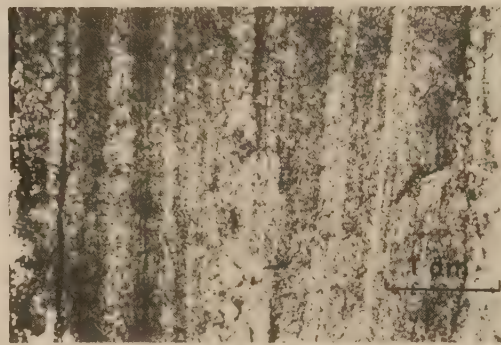


Fig. 26. Banded gneiss with secondary parallel pegmatoid veins (middle and right part). (604/335).

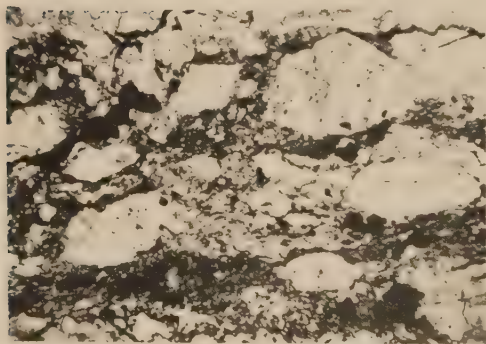


Fig. 27. Granoblastic texture in gneiss. Interlocking of quartz and feldspar. 1 mm. 10x (606/308).

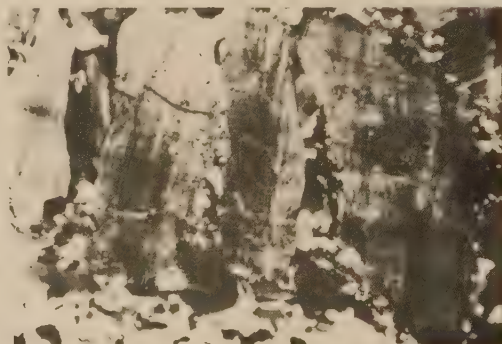


Fig. 28. Granitoid with ruptured alkali feldspar megacrysts. Note the inclusion of sericitized megacrysts (middle, right part). 2 mm. 10x (606/308).

Banding and lamination

These structures are essentially caused by rapid changes in mica contents and are always parallel with the foliation. Due to the sparse outcropping and the often gradual variations between different banded or laminated beds, it is very difficult to follow the separate units in their spatial extension. Three main types of banding or laminations have been observed.

The darker bands at locality 583/346 have a large lateral extension with a maximum width of 15 cm. They are interbedded with 2 - 3 m thick lighter bands, which are richer in quartz and microcline. Chemical differences are shown in Table 12, p. 47.

Another type of parallel structure is seen in an outcrop north of Lyckås (620/296). Here the biotite-richer parts form discontinuous lensoids (Fig. 25), which have a maximum thickness and length of 1 dm resp. 1 m. Although the lensoids are not assembled in such a way that boudinage could be the only explanation of their form, tectonic activity must be an essential factor and probably coincident with the time of foliation. Thus they may primary represent layers of limited lateral extension, which were later tectonically disrupted into lensoids. A metasomatic origin appears less probable.

A third type (Fig. 26) of banding is found NW of Lake Bären (604/335). In this outcrop the biotite-banded gneiss is tending towards homogeneity in the distribution of the minerals. The banding or rather lamination is diffuse, the units are only a few mm thick. Subparallel or parallel with the lamination there may appear coarser-grained neosomes, which are probably initiated by an adjacent cutting pegmatoid.

Within the gneiss area, the banding and lamination seem all over to be parallel with foliation and to be dominated by a north-norhtwesterly orientation. This circumstance and the general concordance of the structures and major lithological contacts, possibly suggest that the banding and the lamination reflect a formation controlled by supracrustal stratification rather than a metamorphic origin. General conclusions of this kind appear to be warranted on the basis that no strictly magmatic rocks

Table 11. Modal analyses (volume percent) of different gneissic rock-types from the Västana area.

Analysis No.	26	27	28	29	30
UTM co-ordinates	605/308	620/296	614/320	604/335	623/296
Quartz	37	38	35	31	20
Plagioclase ¹⁾	30	33	21	25	30
K-feldspar ²⁾	27	22	34	32	31
Muscovite ³⁾	1	+	7	5	13
Biotite	1	4	+	4	5
Chlorite	+	-	-	+	-
Epidote	+	1	1	1	+
Garnet	+	+	-	+	-
Opakes	1	1	1	1	+
Apatite	+	+	-	+	+
Zircon	-	-	+	-	+
Sphene ⁴⁾	+	+	+	+	+
An mol. p.c.					
in plagioclase	2-10	18-25	2-9	5-10	5-10
Average grain-size (mm)	0.14	0.2	0.04	0.1	0.06

+ Indicates accessories (i.e. less than 0.5 % of the mode).

1. Including secondary epidote + sericite

2. " perthitic k-feldspar

3. " sericite

4. " leucocoxene

No. 26. Gneiss, pale red, fine-grained.

" 27. " , pinkish grey, fine-grained, banded.

" 28. " , greyish orange pink, very fine-grained, banded.

" 29. " , light grey, fine-grained, indistinctly banded.

" 30. " , pale yellowish brown, very fine-grained, banded.

Table 12. Chemical composition of some gneissic rock-types from the Västana area.

Analyses No.	26	31	32	29	30	33
UTM co-ordinates	605/308	583/346	614/321	604/335	623/296	583/346
SiO ₂	78.2	76.7	75.6	72.0	69.2	68.4
TiO ₂	0.13	0.10	0.13	0.35	0.38	0.42
Al ₂ O ₃	11.8	12.5	12.7	13.6	15.7	15.9
Fe ₂ O ₃	1.12	0.51	0.79	1.21	1.50	1.62
FeO	0.34	0.54	0.78	1.01	1.30	1.77
MnO	0.03	0.05	0.07	0.11	0.10	0.14
MgO	0.09	0.08	0.39	0.37	0.31	0.93
CaO	0.30	0.33	0.73	1.30	1.20	1.55
Na ₂ O	3.49	3.81	3.19	4.62	4.20	5.55
K ₂ O	4.27	4.82	5.35	3.79	4.50	3.53
LOI	0.33	0.25	0.43	0.42	0.50	0.80
BaO	0.04	0.03	0.06	0.13	0.24	0.08
Sum	100.14	99.72	100.22	98.91	99.13	100.69

No. 26. Gneiss, pinkish grey, fine-grained.

" 31. Quartz-feldspar-rich band in a pale reddish brown, fine-grained gneiss.

" 32. Gneiss, greyish orange pink, very fine-grained, banded.

" 29. " , light grey, fine-grained, indistinctly banded

" 30. " , pale yellowish brown, very fine-grained, banded.

" 33. Dark biotite-rich band in a pale reddish brown, fine-grained gneiss.

Table 13. Niggli values of some gneissic rock-types from the Västana area.

si	533	495	453	384	340	299
qz	+266	+220	+197	+137	+104	+65
al	47	48	45	43	46	41
fm	9	6	11	13	14	18
c	2	2	5	8	7	7
alk	42	44	39	37	34	33
mg	0.10	0.11	0.30	0.22	0.16	0.32
k	0.44	0.45	0.52	0.35	0.41	0.29
ti	0.7	0.5	0.6	1.4	1.4	1.4
h	7	5	9	7	8	12
w	0.74	0.45	0.47	0.51	0.50	0.45
T	+6	+4	+6	+6	+11	+8
t	+3	+2	+1	-2	+5	0

other than those of supracrustal origin, are likely to exhibit structures so well defined and similarly characterized as those of the banded and laminated gneisses.

On the grounds of their stratigraphic position and their textural and chemical similarity with the other banded gneissic rocks the non-banded or non-laminated rock-types (Fig. 24) have been regarded as members of the same formation. Moreover sharp contacts between the rock-types are entirely lacking.

Chemical characteristics

From field evidence and textural traits it is obvious, that although completely recrystallized, the differing grain-sizes in the separate units most certainly reflect a primary feature. This fact suggest a rather long transportation and accordingly fairly well-sorted detritus. In contrast it seems unlikely that thick water-transported deposits of fine-to very fine-grained character should have such a high degree of feldspar survival.

If the banded or laminated gneisses represent epiclastic material and the high alkali content is due to later events, the percentage of the phyllosilicates and the Niggli T and t (Table 13) values are remarkably low. Besides, the whole sequence, which is more than 4 km thick, is too homogeneous to be a sediment transported by water because schists, slates and conglomerates seem to be entirely absent. It must thus be concluded that there is such a high discrepancy between structural and chemical characteristics that an epiclastic origin of the gneisses is not probable.

As already mentioned, the chemical analyses agree very well with the compositions of alkali rhyolites and rhyodacites. Moreover, published analyses (e.g. Larsson 1932, Geijer and Magnusson 1944) of rocks regarded as instances of volcanics from the Leptite Formation in the middle of Sweden show good coincidence.

Several attempts have been made to divide gneisses into ortho- and paragneisses, using chemical analyses plotted in scatter diagrams (e.g. Köhler and Raaz 1951, Leake 1963). However, they are all of no use when dealing with quartz-feldspar gneisses in the Västana area, because the igneous and sedimentary fields in the

binary diagrams most often more or less cover each other. A more congenial approach has been made by D.M. Shaw (1972), who has combined the most useful parameters from the different diagrams into multivariate analyses. His work resulted in the following function:

$$DF = 10.44 - 0.2 \text{ SiO}_2 - 0.32 \text{ Fe}_2\text{O}_3 \text{ (total Fe)} - 0.98 \text{ Mg} + 0.55 \text{ CaO} + 1.46 \text{ Na}_2\text{O} + 0.54 \text{ K}_2\text{O}$$

Positive values of DF indicate probable igneous origin and negative values sedimentary origin. Applied to the chemical analyses in Table 12, they all give positive values ranging from 1.9 to 4.9. Thus in conclusion: There are many indications pointing toward an igneous origin of the gneisses, but due to the sparse outcropping, epiclasts may be present within the formation. Anyhow, on the basis of the present material it is most probable that the overwhelming part of the above-discussed rock-mass is of extrusive lava or pyroclastic origin.

Granitoid rocks

Granitoid rocks crop out in the south-western and north-eastern part of the area. The south-western massif consists of a moderately reddish brown to pale red or reddish grey medium-to coarse-grained usually feldspar-porphyritic granitoid.

A foliation is not always present, but seems to be rather well-marked in the margins of the massif.

Marginal xenoliths of metabasites and metasediments demonstrate the intrusive behaviour of the granitoid. The contacts between the xenoliths and the granitoid are always sharp.

Normally the feldspar megacrysts, which often are poikilitic with inclusions of quartz, are somewhat rounded with a maximum size of 3 cm. When tabular, the megacrysts most often have a preferred orientation and are thus aligned to the semi-penetrative foliation. Frequently, larger crystals are marginally granulated or divided into parts of slightly different optical orientation with fracture fillings of quartz, sericite, and epidote (Fig.28, p.44).

The perthite is usually of patch- or string-type, the former carrying remnants of sericitized plagioclase.

The plagioclase (albite-oligoclase) is strongly sericitized. Grains in contact with alkali feldspar have often clear rims.

The presence of bulbous myrmekite at the margins of the megacrysts in the deformed rocks may be explained by the distribution of pressure around the megacrysts undergoing shearing during the deformation. This shearing may not only have granulated the plagioclase crystal borders, but may also have increased the grain boundary energy, providing a more attractive site for myrmekite crystallization (Phillips & Carr 1973). Inclusions of myrmekite in the plagioclase megacrysts indicate an exsolution genesis of this myrmekite.

Coarser grained quartz is strongly undulous and marginally granulated. The polygonized fine-grained quartz is not undulose and forms a granoblastic texture.

Chloritized biotite and muscovite appear as subordinate minerals. The former has inclusions of sagenite or/and leucoxene. Accessories are opaque minerals, epidote, sphene, chlorite (penninite), apatite, zircon, and fluorite. Allanite cores in epidote are not uncommon.

From the above mentioned it is obvious that the texture indicates cataclasis and a probably contemporaneous recrystallization. Thus it can be stated that at least the megacrysts containing remnants of plagioclase are porphyroblasts, which obviously are pre-tectonic in relation to the foliation and the cataclasis. It is not clear whether the megacrysts should be referred to late-magmatic stages succeeding the original crystallization or to late metasomatic activities connected with metamorphism.

The north-eastern massif is very homogeneous and consists of a greyish orange-pink, medium-grained porphyritic granitoid. The alkali feldspar megacrysts are never larger than 2 cm in diameter and are often perthitic.

Myrmekite is usual in the rims of plagioclase, which often contains a sericitized core. Coarser grained quartz is undulous and fractured. Beside the alkali feldspar, plagioclase and quartz appear as subordinate minerals. Accessories are biotite (partly

chloritized), muscovite, epidote, zoisite, apatite, sphene (leucoxene), opaque minerals, and iron oxide.

In textural respect this latter granitoid is very much like the former type mentioned above.

Dolerite

Dolerite is the youngest rock in the area and appears in dikes. Their dip is always steep and the strike is predominantly north-south. The maximum widths have been estimated to be about 25 m.

The dolerites are medium- to fine-grained, the colour being medium grey to black. The texture varies between ophitic and subophitic with a preponderance for the latter. Plagioclase forms lath-shaped crystals and is the main mineral. Normally, the plagioclase has the composition of basic andesine, but zoning may be present around cores of labradorite. Saussuritization is frequent.

Olivine is an essential mineral and is sometimes marginally pseudomorphosed by serpentine and/or iddingsite. Augite is not as frequent as olivine and may contain chlorite produced by alteration.

Subordinate minerals and accessories are biotite, opaque minerals, apatite, and alkali feldspar.

5. STRUCTURAL HISTORY

Introduction

The map (see inset) shows the outlines of the major structural features in the Västana area. The structures present and the multi-phase train of events are presented in a historical sequence. Each phase of deformation is referred to as D_n , n being the position in the sequence. F_n refers to folds, S_n to penetrative planar fabrics, mainly foliation, and L_n to penetrative linear fabrics.

In general terms, the present attitude of the foliation in the area has a form much like a monoclinial synform whose axial trace

trends NNW-SSE. Later events have folded the penetrative S_1 foliation and the bedding planes S_0 in the metasediments north-west of Lake Bären. Abundant small structures are directly related to these secondary folds, which have a close fold style and a geometry resembling class 1 c according to Ramsay (1967, p. 365).

Later structural events have caused open flexural folding and semi-penetrative to penetrative foliations or have been reflected by structures developed in a brittle state.

D_1 : Foliation S_1 .

D_2 : Foliation S_2 , similar folds F_2 .

D_3 : Foliation S_3 , open flexural folds F_3 .

D_4 : Brittle structures.

Planar structures

Planar structures are well developed in almost all exposures in the area. With the exception of bedding, all planar structures identified and mapped in the field are of tectonic origin.

Bedding

Compositional layering is the planar feature most commonly designated as bedding in metamorphic rocks. The presence of sedimentary rocks substantiates the fact that alternating laminae or layers of contrasting composition represent bedding planes or sedimentary units. In the Västana area the layers and laminae described earlier (p. 36) are bedding planes (S_0).

Foliation plane S_1

The dominant foliation (S_1) is expressed by a preferred orientation of minerals like mica and amphibole. The mica orientation in the metasediments is generally parallel with the lithologic layering (S_0). Even in the fold hinges north-west of the Lake Bären the S_1 foliation is parallel with the bedding planes (Fig. 29B, p. 57) though slightly overprinted by the axial plane foliation S_2 . This underlines the statement above that S_1 is parallel with S_0 .

In the acid gneiss and the metavolcanics the S_1 foliation is a banding of alternating felsic and mafic layers and/or mineral orientation. Whether the compositional banding in the gneisses and the metavolcanics represents true bedding or not is uncertain, but due to the parallelism between the banding and S_0 in the metasediments and the major lithological rock boundaries, this is highly probable.

In the metasediments the foliation (S_1) is defined by orientated biotite and amphibole and by elongated aggregates of quartz and feldspar. Thin streaks, laminae and bands of felsic minerals partly produced by metamorphic segregation are parallel with the foliation.

The mechanisms causing the development of a foliation parallel with bedding have been much disputed in literature. Whitten (1966, p. 258) considered: "It would be desirable for bedding foliation (or cleavage) due to load metamorphism to be expunged as a category". Whitten further states that: "Under least two conditions, foliation can develop parallel to the original bedding".

1. Minerals may grow with preferred orientation along some plane of easy access to solution through the rock, such as selected bedding planes etc.
2. Sometimes flexural-slip folding results in a foliation parallel to the bedding of this incompetent units interlaminated between competent ones.

Thus the mechanism controlling the development of the foliation parallel with bedding can be complex and the processes involved more or less concealed by later metamorphic events.

Foliation plane S_2

In the fold hinges north-west of Lake Bären (Fig. 29) a semi-penetrative foliation is developed parallel or subparallel to the axial planes of the folds. This secondary foliation will be referred to as S_2 . The platy minerals of the early foliation (S_1) are partly crenulated and bent into near parallelism with the

later planes of cleavage (Figs. 31-33, p. 61). These planes are commonly loci of recrystallization.

Foliation plane S_3

Parallel or subparallel the NE-SW trending faults a weak semi-penetrative foliation S_3 has locally been developed within the metasediments. The very weakly orientated mineral growth is mostly only shown on a microscopic scale and seems to be aligned with the axial planes of the F_3 folds.

Foliation plane S_4

The mineral orientation is semi-penetrative and restricted to and parallel with the planes of the north-south trending faults. This connection to structures, which obviously are developed late, suggests, that the designation of the foliation as S_4 is relevant.

Linear structures

Linear structural elements are most prominent within the metasediments. The time relation between the elements are very difficult to distinguish when not directly related to minor fold axes or intersections of S-surfaces. Thus, much of the L_1 measured structures may in time relation be of L_2 character, which is sustained by the nearness between the L_1 and L_2 maxima in the stereograms C, D and E, Fig. 30, p. 58.

Minor fold axes

North-west of Lake Bären and in the hinges of the major F_2 folds there have developed crenulations of the S_0/S_1 planes, box folds, and conjugate kink-banding (Figs. 32 and 33). These minor structures are undoubtedly related to the major F_2 folds and thus seem to be lacking in other regions.

Mullions

In shape, the mullions are long cylindrical structures with rounded and mica-coated surfaces. These structures are parallel with the fold axes and most common in the micaceous quartzite and mica-schist. Sporadic observations of mullion-like structures are found in the quartz-feldspar-gneiss.

Intersections of S-surfaces

This type of lineation is developed as a fine striping, which is parallel to the axes of the F_2 folds, and occurs where the axial-plane cleavage (S_2) intersects S_0/S_1 -surfaces.

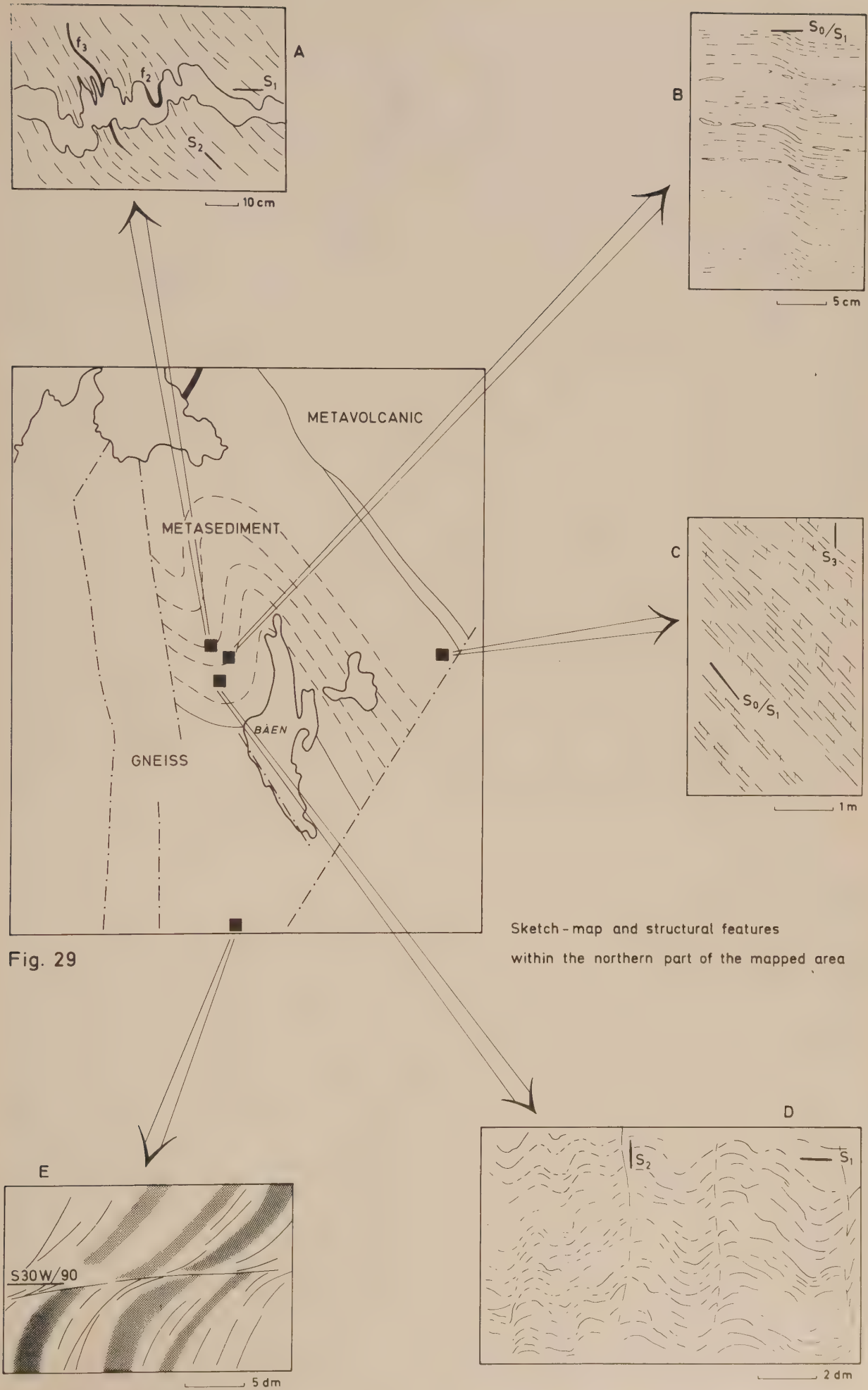
Elongated quartz pebbles

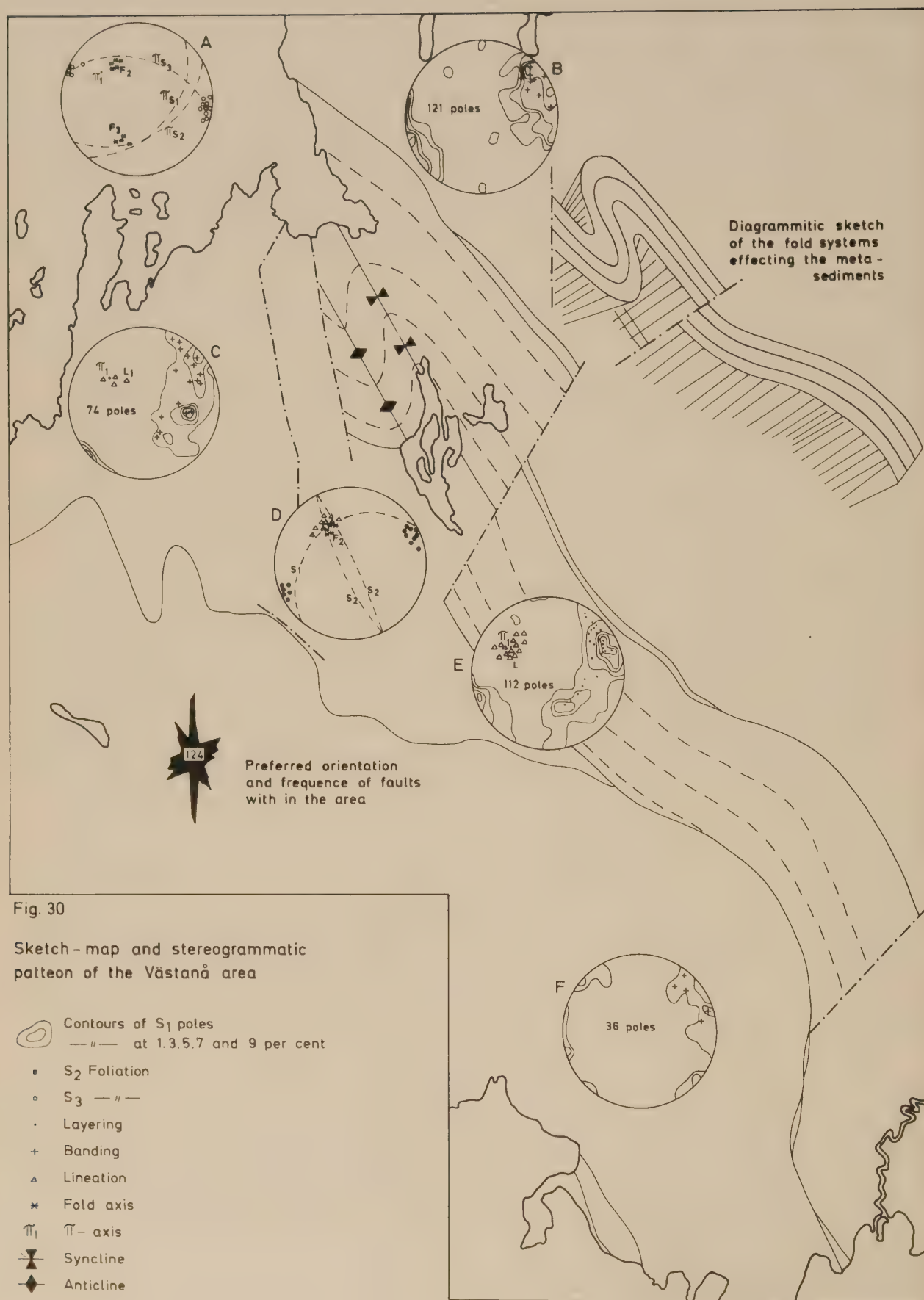
In the intraformational conglomerate the quartz pebbles (p. 30) are flattered and elongated to parallelism with S_0/S_1 . The conglomerate, only exposed in the southern part of the metasediments, has later been very weakly folded by F_3 in a very open flexural fold style (see map). With the above-mentioned reservation (the relation between L_1 and L_2) these lineaments have been regarded as L_1 structures.

Boudinage

Boudinage structures have been developed in the area, where more competent thin quartzites lying in less competent micaceous quartzites or mica-schists have been deformed. Thus, minor-scale boudinage has been observed in the F_2 fold hinges (Fig. 29 B).

In detail, the boudins are somewhat irregular in shape and their distribution within a layer is usually unsystematic. The boudins are generally strongly elongated and separated and have normally widths of 1 - 2 cm and lengths of 3 - 5 cm.





D₁-deformation

The D₁ structures found within the area have no corresponding minor or visible major folds to which they can be related. Both structural readjustment and mineral constitution of the dominant fabric of the rocks took place during this phase of deformation and are well displayed in the foliated and banded rock types. None of the following phases of deformation (D₂ to D₄) shows such a penetrative mineral alignment. This indicates a relatively high intensity of tectonic deformation during the D₁ phase.

By plotting the orientation of the planar and linear structures on equal-area lower hemisphere stereograms, it is possible to demonstrate the degree of preferred orientation of various structural features. First-deformation structural elements are illustrated in Fig. 30, stereograms A to F.

The S₁ foliation is recognizable throughout the whole area, except where it has been obliterated by extreme attenuation during D₂ and D₄. When regarding the rather uniform distribution of outcrops within the region and the high degree of preferred orientation of the S₁ poles (Fig. 30, stereograms B, C, E and F in the northeastern quadrant), the S₁ foliation is similar to those planar structures which are expected to be found in folds with well developed limbs.

The lineation designated as L₁ in the metasediments expressed by elongated quartz pebbles (Fig. 30, stereogram C), which have been described on p. 30. In the gneiss the L₁ lineation is represented by a rather fine striation or corrugated surfaces. The orientations of the lineation L₁ developed in the gneisses and the metasediments show good correspondence and are not visibly affected by any later deformation.

If the S₁ foliation and the L₁ lineation are structures developed during large-scale folding, then the fold axis should have orientations which are close to the constructed pole (Fig. 30, stereograms C and E) with which the lineaments (L₁) also show good agreement.

D₂-deformation

As has been mentioned, the S_2 foliation is best exposed in the fold hinges of the major folds north-west of Lake Bären (Fig. 29 A). These F_2 folds have a close fold style and are only developed in the less competent part of the metasediments within rocks such as mica-schist and micaceous quartzite.

The hinges of the major F_2 folds are crenulated and kink-banded and the axial planes of the small crinkle-folds in the micaceous layers produce a parallel structure that crosses the earlier foliation S_1 and the layering S_0 .

In rocks affected by crenulation, the neocrystallization has been non-penetrative and thus restricted to actual planes of slip. These planes of slip most often occur within the limbs of small-scale folds, where the mica flakes are bent into parallelism with the axial planes. The axial planes of the major F_2 folds and the small scale folds seem to coincide and the axial plane foliation of the major folds (S_2) and the crenulation foliation are thus parallel. Although the intensity of the S_2 foliation may vary, it never becomes as penetrative as S_1 . This variation in intensity of S_2 within the major F_2 folds (Figs. 29 A and D), may depend on the difference in competence between the layers or may be a reflection of the internal distortion during the processes of folding.

On a minor scale the deformation of the pre-existing planar fabric is shown in the hinges of the major F_2 folds by the folding of muscovite flakes in micaceous layers (Figs. 32 and 33). The micaceous minerals in the rock are folded and forced into parallel alignment, which in this case is parallel to S_2 . As the intensity of folding increases the small-scale fold hinges are depleted of muscovite, which seems to concentrate in the fold flanks. The multifariousness of the structures is represented by box-folds, conjugate folds, crenulation, and single and conjugate kink-bands. Thus the fold styles are heterogeneous and transitions between the different types are common. From Fig. 32 it is obvious that this single fold is both a box-fold and a conjugate fold.



Fig. 31. Crenulated and kink-banded mica-schist. Strain-slip cleavage along S_2 . (607/337).

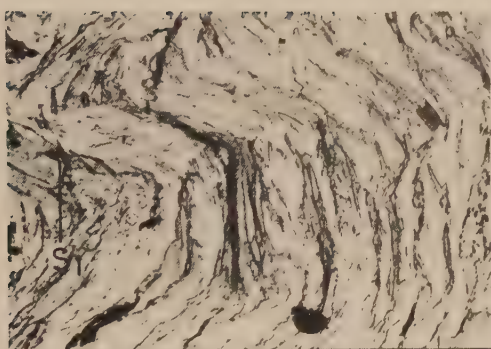


Fig. 32. Micro-folding in a mica-schist. Successive stages between conjugate folds (left) and box-folds (centre and right). 1 nic., 16x (608/337).



Fig. 33. Crenulated mica-schist with S_2 foliation weakly developed along the limbs of the fold. 1 nic., 16x (608/337).

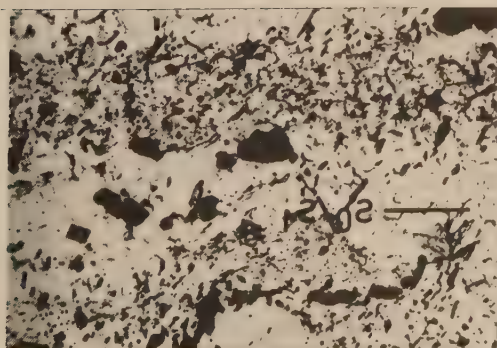


Fig. 34. Thin layers of iron oxide parallel with S_0/S_1 . 1 nic., 16x (609/335).

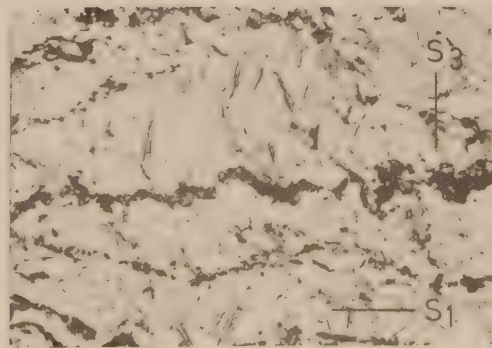


Fig. 35. Tiny crystals of kyanite and chloritoid parallel with S_1 . The S_2 foliation is marked by oblique orientated muscovite flakes. 1 nic., 16x (611/336).

The successive surfaces in the left part of the photo are nearly identical in form, which according to Ramsay (1967, p. 358) characterizes conjugate folds. However, in the extreme right part of the photo the polyclinal shapes change into a single-hinge type of folds, which are distinctive for box-folds.

Kink-banding is rather frequent and can be separated into single and conjugate kink-banding. The former is developed as angular, steplike monoclines formed on parallel-plane axial surfaces and the latter consists of a polyclinal structure described as conjugate folds. The obviously very close relationship between the different fold styles and kink-banding suggests that these structures may be contemporaneously developed.

Boudinage is another small-scale structure present in the hinges of the major F_2 folds. Thin competent quartzitic layers show pinch and swell structures or have partly been split up into somewhat elongated pieces with thinned necks (Fig. 29 B).

The change of S_1 from a northwesterly strike to northnortheast as marked by the polar submaxima in the southeast quadrant (Fig. 30, stereogram E), is due to folding of S_1 . The fold axes (F_2) are congruent and lie in the axial plane S_2 . Though the fold axes and lineations (L_2) are rather tightly grouped (Fig. 29, stereogram D), the fold axes seem on average to have a slightly steeper plunge.

With common axial planes and orientations of the lineation, it is clear that the major and small-scale structures are closely connected.

The mechanisms for the evolution of conjugate folding and kink-banding have been much discussed in the literature. Ramsay (1967, p. 438) considered that: "The internal deformation that goes on within the material folded by these structures appears to be almost completely accomplished by slip on the layering as with the flexural-slip model, and the surface of the layers may show signs of slickensiding. The folds form as a response to a compressive stress acting along the layering.....".

Cobbold et. al. (1971, p. 51) stated that: "Sinusoidal buckles (crenulation) and conjugate kink-bands are not isolated structural

types formed by different mechanisms, but are the end members of a range of possible fold structures.... The idea, for example, that kink-bands develop in "brittle" environments only is erroneous."

Though there are some deviating opinions, most of the authors agree in that these structures are caused by compressive stress acting along the layering. The folds result from relative displacement in a direction parallel to their axial planes i.e., normal to the compression. If the small-scale structures are indicative of the processes involved during the development of the major F_2 folds, these are probably formed by buckling. If the rock materials change their properties during the deformation, then faults and slips may develop along the axial surfaces and lead to rupture of the kink-zone. (Fig. 31). Thus, although the axial surfaces are not primary shear planes generating the folds, they are commonly taken over by faulting at a later stage of fold development. The shear planes that are developed in this way by deformation of foliation are often termed strain-slip cleavage (Ramsay 1967, p. 388). The kink planes form along the planes of high shear and are positions of restricted cataclasis, which may have caused the partial granulation of quartz and the kinking of mica and kyanite.

D₃ - deformation

The structures developed in the D₃ phase are folds and a connected semipenetrative foliation (Fig. 29 C). The fold style is open with SSW-trending and moderately plunging fold axes. On a larger scale the very open fold style is sustained by the major flexure north of Klagstorp. Here the nearly uniform thickness of the lithological units (S_0) and the very weak development of the axial plane foliation show the concentric nature of the folds of the D₃ phase.

Lineations seem to be entirely missing.

The S_3 foliation is only locally developed by the subparallel alignment of thin muscovite flakes. This semipenetrative foliation

is most often shown only on a microscopic scale by mica orientated at high angles to the dominant foliation S_1 or S_2 (Fig. 35). In the stereogram A, Fig. 30 both the plot of the fold axes and the poles of the S_3 foliation show rather good concentration.

This deformation (D_3) seems to be very restricted in appearance and importance. Like the D_2 phase, the D_3 deformation has caused minor cataclasis of feldspar and quartz and kinking of mica and kyanite.

D_4 -deformation

This deformation is developed essentially during a brittle stage. Faults are rather common within the area and generally have vertical dips and northerly, northwesterly or northeasterly strikes. The previously deformed rocks are thus cut by faults, which cause disruption and deformation in narrow fault zones. The preferred orientation and frequency of the faults in the area can be seen in Fig. 30.

The NNE-SSW trending faults seem to be, at least partly, combined with strike-slip displacements, which show a maximum offset of 1 km. From the fault crossing the metasediments just south of N. Grönhult, it is obvious that the displacement also included a vertical movement. Fig. 29 E shows a NNE-trending fault and the relative direction of the slip. In the very boundary of the slip the banding has been forced into a structure that is more or less subparallel to the orientation of the fault.

The NNW-SSE trending faults are of minor significance. These faults are especially frequent near the margins of the granitoids in the vicinity of Lake Bären.

The most important fault set runs in a north-southerly direction. High-angle faults are most common and the slickensides indicate dip-slip movement. These lineaments are easily traced on the topographical map and are reflected by the outlines of the lakes. Thin zones of mylonite and a semipenetrative foliation (S_4) are often connected with the displacements. In the

gneiss area north-west of Lake Bären seismic investigations have been carried out. The obtained data have confirmed the presence of N-S trending anomalies of much lower wave velocities than those in the gneiss. These crushing- or mylonite zones have a estimated maximum width of 50 m and indicate a rather thorough tectonization of the rock.

Southwest and west of Lake Bären low-angle ($25-35^{\circ}$) faults have been observed within the quartz-feldspar-gneisses. This type of N-S trending faulting diverges from the normal high-angle faults in the other rock types and indicates that the major principal stress was essentially horizontal.

Tentative structural development

The structures in the Västana area indicate that the rocks have suffered at least two periods of folding (D_2 and D_3) during which there developed two different fold systems accompanied by related minor structures.

As mentioned earlier, the D_1 deformations have no corresponding minor or visible major folds to which they can be related. The S_1 foliation is well accentuated and seems to be the most penetrative structure within the mapped area. Outside the F_2 folds S_1 seems to coincide both with the orientation of S_0 and S_2 . The latter is obviously not as penetrative as S_1 , which is well demonstrated by the appearance of S_2 in the hinges of the F_2 folds (Fig. 29 D). The S_1 foliation may have developed in the limbs of a large isoclinal fold, the further extension of which has been obliterated by later deformation. If so, the fold axis of the large scale fold should be parallel with the L_1 lineation and thus trending NNW with a moderate plunge. This suggests that the metasediments and associated rock types are situated in the eastern limb of a large synclinal structure.

During the D_2 deformation S_0 and S_1 were folded and gave rise to abundant small structures in the hinges of the major F_2 folds. The large-scale folds have a close fold style whereas the style of minor folding is most variable.

The time relation between D_1 and D_2 is uncertain, but it is not impossible that they are closely connected. This is sustained by the almost parallel orientation of S_1 and S_2 as well as L_1 and L_2 , which indicates that D_1 and D_2 were developed under similar physical conditions. Moreover, the D_2 phase has only been found in less competent units such as micaceous quartzite and mica-schist, which suggest that the major F_2 folds may be regarded as parasitic folds, developed in the limb of a large scale fold.

The D_3 deformation is weak and shows an open flexural fold style, which is associated with a semipenetrative axial-plane foliation. Parallel to the curvature of the large-scale flexure north of Klagstorp follows the contact between the granitoid and the metasediments. This and the earlier described intrusive character of the granitoid (p. 50) indicates that the D_3 deformation may have been connected with the granitoid intrusion.

The D_4 deformation has been developed during a brittle stage and is exclusively associated with the final phases of deformation. All the rock types within the area are affected by this phase, which also has caused mylonitization and some neocrystallization.

6. TIME RELATION OF DEFORMATION AND MINERAL GROWTH

In this section it is proposed to examine the metamorphic history in relation to the polyphase deformation. It is clear that the Västana region has undergone at least four episodes of deformation during which there occurred crystallization of metamorphic minerals. However, due to the lack of evidence, the correlation of definite minerals with special deformation events has been impossible and accordingly this chapter is somewhat tentative.

Crystallization and deformation are not necessarily simultaneous and may begin and end at different times. The crystallization of a mineral may be designated pre-, syn-, or post-tectonic, according to whether it crystallized before, during or after a given phase of deformation.

METAVOLCANIC FORMATION

	D ₁			D ₂		D ₃		D ₄	
QUARTZ			---						
PLAGIOCLASE									
BIOTITE						---			
MUSCOVITE						---			
CHLORITE						---			
AMPHIBOLE				---					
GARNET		---		---					
DEFORMATION STAGE	PRE	SYN	POST	SYN	POST	SYN	POST	SYN	POST
	TECTONIC			TECTONIC		TECTONIC		TECTONIC	
S PRODUCED	S ₁			S ₂ ?		S ₃		S ₄	
S FOLDED	S ₀ ?					S ₀ , S ₁			

METASEDIMENTARY FORMATION

	D ₁			D ₂		D ₃		D ₄	
QUARTZ			---		---		---		
MUSCOVITE							---	---	
CHLORITOID				---				---	---
KYANITE									
ANDALUSITE		---		---					
OPAQUES									
DEFORMATION STAGE	PRE	SYN	POST	SYN	POST	SYN	POST	SYN	POST
	TECTONIC			TECTONIC		TECTONIC		TECTONIC	
S PRODUSED	S ₁			S ₂		S ₃			
S FOLDED	S ₀ ?			S ₀ , S ₁		S ₀ , S ₁ , S ₂			

GNEISS COMPLEX

	D ₁			D ₂		D ₃		D ₄	
QUARTZ				---		---			
PLAGIOCLASE									
BIOTITE				---	?				
MUSCOVITE				---	?	?			
CHLORITE									
GARNET				---	?				
DEFORMATION STAGE	PRE	SYN	POST	SYN	POST	SYN	POST	SYN	POST
	TECTONIC			TECTONIC		TECTONIC		TECTONIC	
S PRODUCED	S ₁			S ₂ ?		S ₃ ?		S ₄	
S FOLDED	S ₀ ?								

Fig. 36 Diagram showing the relation between mineral growth and deformation

Metavolcanic formation

In the acid and intermediate metavolcanics the platy minerals define a penetrative foliation (S_1) and consequently are syn-tectonic with D_1 . The pre-tectonic crystals, such as phenocrysts which have not been obliterated during the D_1 phase, are commonly somewhat modified, granulated or fractured and are enveloped by orientated mica flakes. This envelopment texture has been interpreted by various writers, among them Zwart (1960, p. 215), who considers that the large crystals may be post-tectonic and the micas swirling around them had been pushed aside as the megacrysts grew. However, the presence of granulated crystals and pressure shadows indicates that these phenocrysts are indeed not post-tectonic, but pre-tectonic as regards D_1 .

The D_2 -phase could not be traced.

Later deformation caused mainly mechanical rotation and decomposition of the biotite, which is sometimes completely transformed into chlorite pseudomorphs. In the tuffites this later structural event also resulted in at least a partial parallel alignment and recrystallization of muscovite, thus forming a weak S_3 foliation.

Undeformed poikilitic and random orientated muscovite flakes indicate a post- D_3 inter-kinematic growth.

The D_4 phase caused the recrystallization of quartz and muscovite and the neocrystallization of chlorite.

In the metabasites the relation between deformation and mineral growth is more controversial. Biotite flakes and tiny amphibole prisms show parallel alignment with S_1 . This implies a syn- D_1 crystallization.

Thin secondary quartz-veins, which cut the S_1 foliation are pierced by larger amphibole crystals, which sometimes are in parallel alignment with S_1 . Other amphibole grains are orientated randomly or are aggregated into "rosettes" and consequently cut the S_1 foliation. This suggests that although most of the secondary amphibole crystals seem to have been formed

during an inter-kinematic stage, which post-dates syn-D₁, they must also partially have crystallized either during a syn-D₂ stage or have grown mimetically along existing S-planes.

More controversial is the growth of garnet. The porphyroblasts are subhedral or euhedral and often contain irregular cracks, which are healed by quartz. Garnet is always connected with a thorough epidotization or with secondary salic veins forming stripes or bands parallel with S₁. It is obvious that these metamorphic segregations have developed along earlier S-planes, which postulates a crystallization later than D₁. Because the garnet porphyroblasts cannot be related to any of the foliations and do not contain tell-tale internal trails or inclusions, the relation between deformation and garnet growth is obscure.

However, because the D₃-phase is usually associated with only a weak chloritization of biotite and thus only has not caused more than a limited recrystallization or neocrystallization and orientation of the micas, it is suggested that the amphibole and garnet crystals were formed together during a stage earlier than D₃.

It appears that the D₄ deformation is not present in the metabasites.

Metasedimentary formation

The relationship between the foliations S₁, S₂, and S₃ in the metasediments is shown in Figs. 31-35, p.61. All these foliations, which are marked by muscovite, are attributed to movements during the regional metamorphism and are accompanied by the growth of metamorphic minerals. Thus it is clear that muscovite has recrystallized syn-tectonically with each of the three deformation stages.

Kyanite occurs in micaceous quartzite or in mica-schist and forms two different types of crystals. Short prismatic or platy subhedral grains tend to have their longest axes orientated parallel with the S₁ foliation. Larger individuals

are orientated randomly and are commonly aggregated. Both types of kyanite are shown in Fig.15, p.33. The very strong alignment of the tiny kyanite prisms along the S_1 -planes rather indicates a syn-kinematic growth than a mimetic one.

Thin rows of iron oxides, which are parallel with S_0/S_1 , have been enclosed in the larger porphyroblasts. Because these inclusions have retained their original orientation, they indicate that the host have not been rotated. Moreover, the larger crystals are obviously not affected by the penetrative foliation S_1 . All this implies that the larger randomly orientated kyanite crystals have grown during a post- D_1 inter-kinematic stage. Bent or fractured kyanite crystals found in the fold hinges of the major F_2 folds, indicate deformation during D_2 and thus suggest that the growth of kyanite ended before the syn- D_2 stage.

Andalusite forms anhedral or subhedral porphyroblasts, which usually carry poikilitical inclusions composed mainly of quartz (Fig. 20). The S_1 foliation has not deformed the andalusite crystals, which indicates a growth after the D_1 stage. Also the andalusite may contain thin rows or clouds of iron oxide grains, which are often concentrated to the central parts of the porphyroblasts.

Inclusions of kyanite in andalusite (Fig. 19), which exhibit no exsolution or replacement relation with their host demonstrate that the andalusite is younger than the enclosed kyanite.

Also, when kyanite and andalusite occur alongside, there is no signs of transformation of one polymorph into the other. Andalusite crystals found in the hinges of the major F_2 folds, are not deformed during D_2 . In contrast, andalusite porphyroblasts in other parts of the metasediments have clearly been rotated and/or deformed by D_3 . This implies that the growth of andalusite most probably began during a post- D_1 stage and ended in a post-kyanite stage, which was clearly pre- D_3 .

Chloritoid is a common constituent in the mica-schists and micaceous quartzites. It coexists with kyanite, but never with

andalusite. As was the case with kyanite, tiny acicular or prismatic chloritoid crystals are assembled into thin lamellae where the individual crystals tend to have their longest axes orientated parallel with the foliation S_1 .

Larger porphyroblasts are most often aggregated and randomly orientated, which suggests crystallization during a inter-kinematic phase. The intimate association of kyanite and chloritoid and the similar behaviour in relation to deformation, indicate that these minerals were largely formed together. However, inclusions of kyanite in chloritoid imply that chloritoid continued to grow after the end of the kyanite crystallization.

Gneiss complex

In the gneiss complex the relations between mineral growth and deformation are ambiguous. Although there is some foliation, it is often vague and only marked by biotite flakes orientated subparallel to S_1 .

Later tectonic events have caused the granulation of the quartz and feldspar megacrysts and have transformed the biotite into semi-opaque pseudomorphs. Quartz and sericite appear to have crystallized in thin "layers" sweeping around the large relic "porphyroclasts". The granulation was developed during the stages of more or less brittle deformation and may be connected with any stage later than D_1 .

Large disseminated poikilitic flakes of muscovite show no signs of deformation and must consequently have crystallized during a late post-kinematic stage. This postulates a post- D_3 growth.

The growth of garnet has predominantly followed the S -surfaces, which are parallel with S_1 and is sometimes connected with a diffuse metamorphic segregation. In this regard the gneiss garnets resemble the garnets of the metabasites. Thus, once more, the relation between garnet growth and deformation is uncertain.

Thin mylonite zones were formed during the latest phase of deformation (D_4), which caused a weak recrystallization of chlorite.

Granitoids

The granitoids have obviously been subject to a widespread cataclastic deformation. It is not known whether this tectonization is to be referred to D_4 or is the result of multiple deformation. However, D_4 is traceable along N-S trending faults where it caused the appearance of very thin tectonized zones with a subparallel alignment of the micas.

7. METAMORPHIC HISTORY

Chemical and modal analyses of characteristic rock-types have demonstrated that there are clearly different groups of formations. Disregarding the igneous rocks, it can be shown that the development of the metamorphic mineral assemblages in the Västana area was mainly controlled by the chemistry of the primary sedimentary and volcanic units. This will be discussed in terms of regional metamorphism. However, the metamorphic evolution of the different rocks is complex and thus it is impossible to separate and establish the exact number of metamorphic events. The inferred metamorphic grade must accordingly be parallelized with the presumed peak of metamorphism, which is still traceable in the mineral associations.

In order to describe the temperatures and pressures of the metamorphic process, the rocks listed in the following tabulation were considered:

Metavolcanics	Metabasite	Micaceous quartzite	Gneiss
Plagioclase (An 5-24 %)	Plagioclase (An 25-40 %)	Quartz	Plagioclase (An 5-10 %)
Quartz	Biotite	Muscovite	Quartz
Biotite	Hornblende	Kyanite	Muscovite
Epidote	Epidote	Andalusite	Biotite
Muscovite	Chlorite	Chloritoid	Epidote
Hornblende	Garnet		Chlorite
Chlorite	Calcite		Garnet

The metavolcanics display an association of acid plagioclase, quartz, biotite, epidote, and hornblende. This corresponds to the higher part of the greenschist facies (Winkler 1967). Going westwards from the acid into the intermediate metavolcanics the plagioclase becomes more calcic and hornblende appears. The transition zone is diffuse. Difficulties arise in assigning these assemblages to metamorphic facies and the mineralogy may correspond either to the upper part of the greenschist facies or to the lower part of the amphibolite facies.

As given in the above list, the mineral association of the metabasites is characteristic of the lower part of the amphibolite facies. Possibly, there may thus occur a slight increase in metamorphic grade when proceeding from the acid metavolcanics into the metabasites.

As far as the metasediments are concerned, the mineral assemblage quartz, muscovite, kyanite, andalusite, and chloritoid suggests that the metamorphic grade corresponds to the lower or middle parts of the amphibolite facies.

The mineral assemblage of the gneiss unit belongs to the upper part of the greenschist facies.

Of the minerals mentioned above, the aluminosilicates and chloritoid have been studied experimentally and may therefore

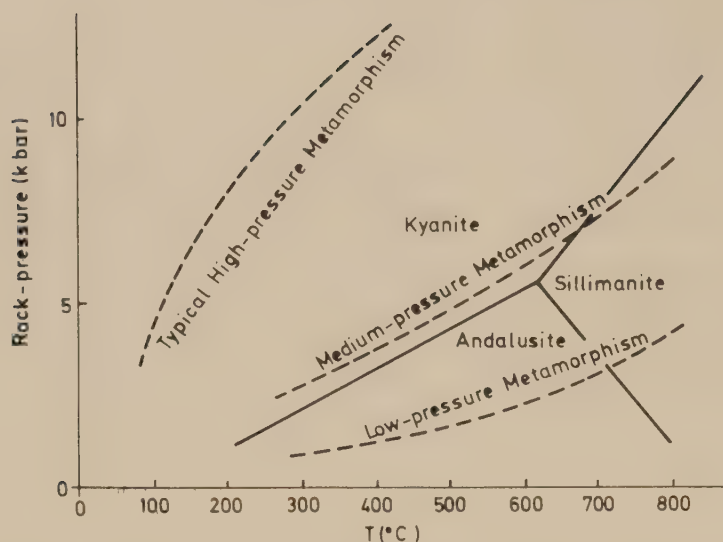


Fig. 37 Classification of metamorphic facies series in relation to the stability fields of Al_2SiO_5 minerals. After Miyashiro (1973).

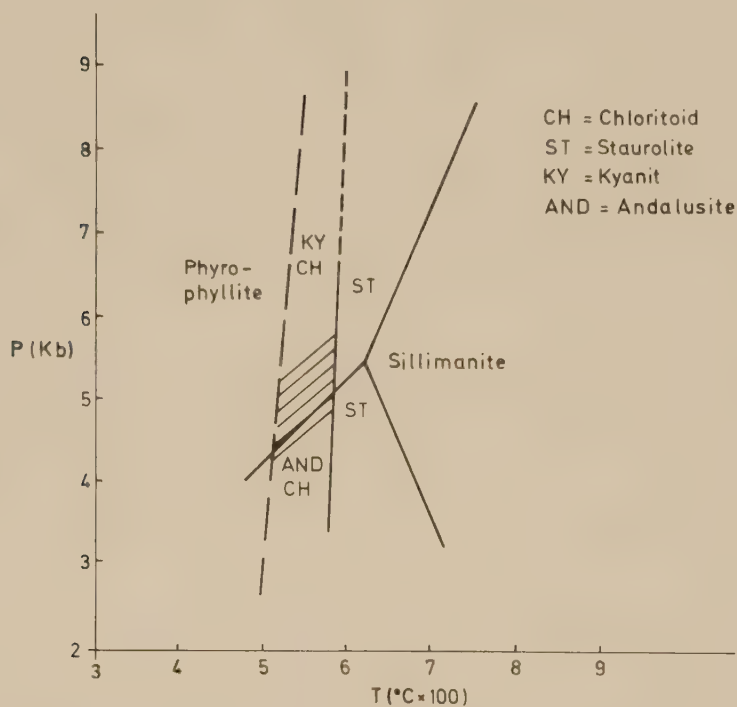


Fig. 38 Pressure-temperature, grid of pertinent reaction curves. Shading marks the presumed limit of the P-T conditions during the regional metamorphism. Equilibria after Hoschek (1969) and Richardson et. al. (1969).

provide information on the P-T conditions prevailing during the regional metamorphism of the region. The dualism between andalusite and kyanite divides the metamorphic facies into Abukuma- and Barrovian-types, and high- and low-pressure facies series (Fig. 37). Of course there are all transitions between these series.

Experimentally determined or calculated P-T fields for the index minerals kyanite, andalusite, and chloritoid are plotted in Fig. 38. This representation seems to provide a reasonable framework, but it may have limitations due to the errors involved, such as the accuracy of experimental determination and variations caused by $P_{\text{tot}} = p_{\text{H}_2\text{O}}$. These problems concern particularly the aluminosilicates and there is at present some disagreement among the different experimental determinations of the aluminosilicate tripple point.

From the petrographical observations it is clear that in the Västana area there is coexistence of kyanite and chloritoid and kyanite and andalusite. The association andalusite and chloritoid does not seem to occur in the area.

Andalusite and kyanite are associated with quartz and must accordingly have been within the stability field of the Al_2SiO_5 - quartz assemblage. The low-temperature limit of the aluminosilicates is defined by the reaction $\text{pyrophyllite} \rightleftharpoons \text{Al}_2\text{SiO}_5 + \text{quartz} + \text{H}_2\text{O}$. Another reaction of interest is the transition $\text{chloritoid} + \text{Al}_2\text{SiO}_5 \rightleftharpoons \text{staurolite} + \text{quartz} + \text{H}_2\text{O}$. This phase boundary should according to Hoschek (1969) be around $545 \pm 20^\circ\text{C}$ at pressures between 4 000 and 8 000 bars. A minimum pressure limit also defined by the mineral assemblage kyanite and chloritoid.

The geographically very close connection between the andalusite-kyanite and kyanite-chloritoid associations indicates that the difference in P-T conditions between the two mineral assemblages probably is very small. Alternatively, the separate mineral associations are a result of a very restricted and uneven retrograde metamorphic development. Anyhow, in the meta-sediments the P-T conditions of regional metamorphism must have

been close to the phase boundary kyanite-andalusite (Fig. 38). Later metamorphic events in the area have caused a widespread retrogressive greenschist facies metamorphism, which can be traced by the alteration of higher-grade minerals.

8. GENERAL CONSIDERATIONS

A generalized pattern of the Precambrian evolution in the Västana area is summerized in the following scheme:

Faulting, cataclasis, local mylonitization
Intrusion of granitoids and pegmatitic veining

Intrusion of basic dikes
Faulting and local mylonitization
Intrusion of the older granitoids in Blekinge
Dike amphibolites
Folding
Formation of the Coastal Gneiss and deposition of
supracrystals of the Västana complex
(Oldest)

The Coastal Gneiss passes over into metavolcanics, which underlie metasediments such as ortho-quartzite, micaceous quartzite, and mica-schist. These sediments are upwards followed by fine-grained gneisses.

As mentioned earlier, the interbedded epiclastic sediments in the metavolcanics suggests a rather even transition from a dominant pyroclastic to an epiclastic deposition. A more significant transition in the geological development is indicated by appearance of the fine-grained red or grey gneisses, which probably are of volcanic origin.

The metamorphic grade within the investigated area varies between upper greenschist and lower to middle amphibolite facies. Regarding the P-T conditions the presence of kyanite, chloritoid, and andalusite are of special interest. Kyanite

has been reported from several localities in western Sweden, all of which seem to be situated within or very close to the Småland Suture Zone. Thus kyanite is described from Dicksberget, Hålsjöberg, and Västana. However, from modern experimental data kyanite does not necessarily indicate regionally high pressures or a very steep P vs. T gradient. The mineral paragenesis in the Västana area do not indicate extremely high-pressure conditions. As mentioned earlier, the presumed peak of metamorphism in the metasediments has probably occurred at medium pressures (4-6 kb) with a temperature ranging between 500 and 600 degrees C.

The general foliation strikes NW or NNW and has a steep westerly dip. Just northwest of the mapped area there outcrop tectonized acid fine- to medium-grained rock-types having N-S or NNE-SSW striking foliation. According to earlier opinions the Småland Suture Zone should appear in the Västana area or its westerly extension.

Thin and very restricted zones of mylonite do exist in the Västana area and if these correspond to the Småland Suture Zone, this zone must here consist of a series of very thin and insignificant faults running in an approximately north-southerly direction.

Going westward from the mapped area it is impossible to find a strict border line of a zone, which could possibly correspond to the mylonites of the Småland Suture Zone. There are only very thin zones of mylonite no broader belts. Thus, within the mapped area and its immediate surroundings there is no large-scale mylonitization, but instead cataclasis deformation, which type of tectonization has affected large areas and thus belongs to the regional tectonic pattern. In conclusion, no zone which could be equalized with the Småland Suture Protogine Zone could be discovered.

. The available isotopic data from the Västana region are restricted to one single K/Ar determination of a mica quartzite, which has rendered an age of 1240 m.y. In the light of the metamorphic events, which have affected the area, this age must be

considered very difficult to interpret.

A Coastal Gneiss from the area northwest of Karlskrona, has given a K/Ar age of 1560 m.y.

A recent Rb/Sr dating of a young granitoid just south of the mapped area shows an age around 1500 m.y. This granitoid has intruded the Västana complex, which must consequently be older.

Due to the lack of radiometric age data, any correlation with other supracrustals complexes must be tentative. The supracrustals in the Västervik and Vetlanda regions may possibly be chronologically equivalents. In Vetlanda and Västana and probably also in Västervik the supracrustals are undoubtedly older than the surrounding granitoids. Consequently, it is possible that the ages of these supracrustals complexes are syn- or even pre-orogenic in regard to the Svecofennian evolution.

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